Atmospheric Science
An Introductory Survey
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Chapter
3
Extratropical Synoptic-Scale Disturbances

Throughout middle and high latitudes, day to day weather changes are closely linked to the passage of transient, synoptic-scale disturbances in the tropospheric wind field. Through the systematic display and analysis of synoptic (that is, simultaneous) surface and upper air observations, such disturbances can be identified and tracked through the course of their life histories. In this chapter we will examine an individual winter storm system using conventional synoptic analysis techniques that reveal its three-dimensional structure and time evolution. It should be emphasized at the outset that no two of these disturbances are exactly alike. It is only through the examination of a large number of systems (for example, by following the current synoptic charts over a period of months) that it is possible to develop some appreciation for the wide range of structures and time sequences that are possible. For this example we have attempted to select a storm system that is reasonably representative, to the extent that any single example can be representative of a class of diverse phenomena.

3.1 THE 500-mb FLOW

Figure 3.1 shows the hemispheric distribution of the height[†] of the 500-mb pressure surface at or just after midnight (00)[‡] Greenwich Civil Time (GCT) 20 November 1964. These charts are constructed from measurements obtained from radiosondes, which are launched simultaneously from hundreds of stations scattered over the hemisphere. We recall from Section 1.6 that the winds tend to blow parallel to these height contours, *leaving low* heights to the *left* in the northern hemisphere. Wind speed tends to be inversely proportional to the spacing between the contours, which are drawn for every 60 m change in height.

[†] Strictly speaking, we refer here to *geopotential height* as defined in Section 2.2.1.

[‡] Unless otherwise noted, times will be expressed in whole hours.

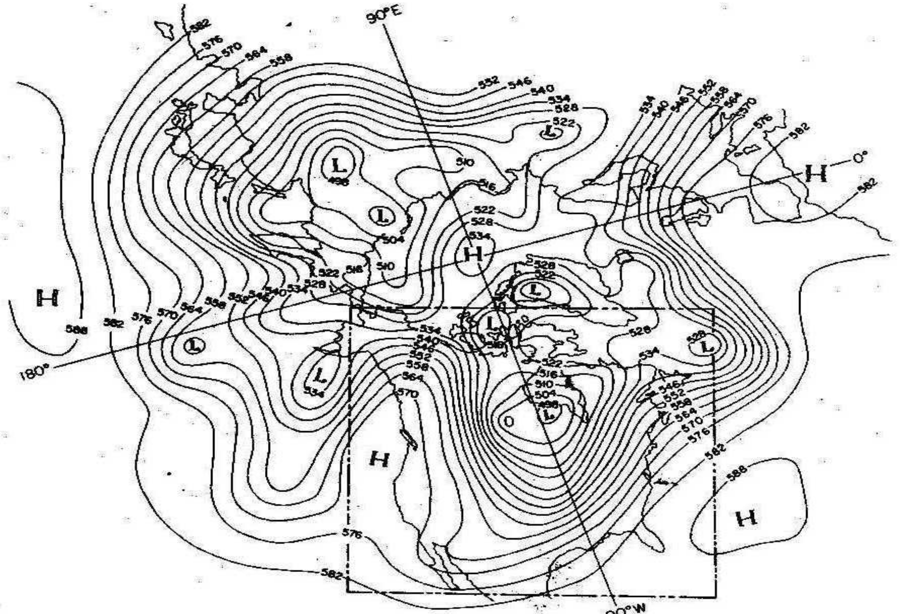


Fig. 3.1 The distribution of geopotential height on the 500-mb surface at 00 GCT 20 November 1964. Labels on contours represent geopotential height, in tens of meters. The letters H and L denote centers of high and low geopotential height, respectively.

At any given instant in time, the hemispheric flow pattern is dominated by large amplitude, synoptic-scale disturbances which vary from day to day in position and intensity. When averaged over a long time period, such as a season, these transient ridges and troughs tend to cancel one another. Hence the climatological 500-mb flow pattern shown in Fig. 1.14 is relatively featureless when compared to a typical instantaneous flow pattern such as the one depicted in Fig. 3.1.

In order to simplify the task of following the time evolution of the 500-mb flow pattern we will concentrate on the limited area enclosed by the rectangle in Fig. 3.1. Four successive 500-mb synoptic charts, spaced at intervals of 12 h,

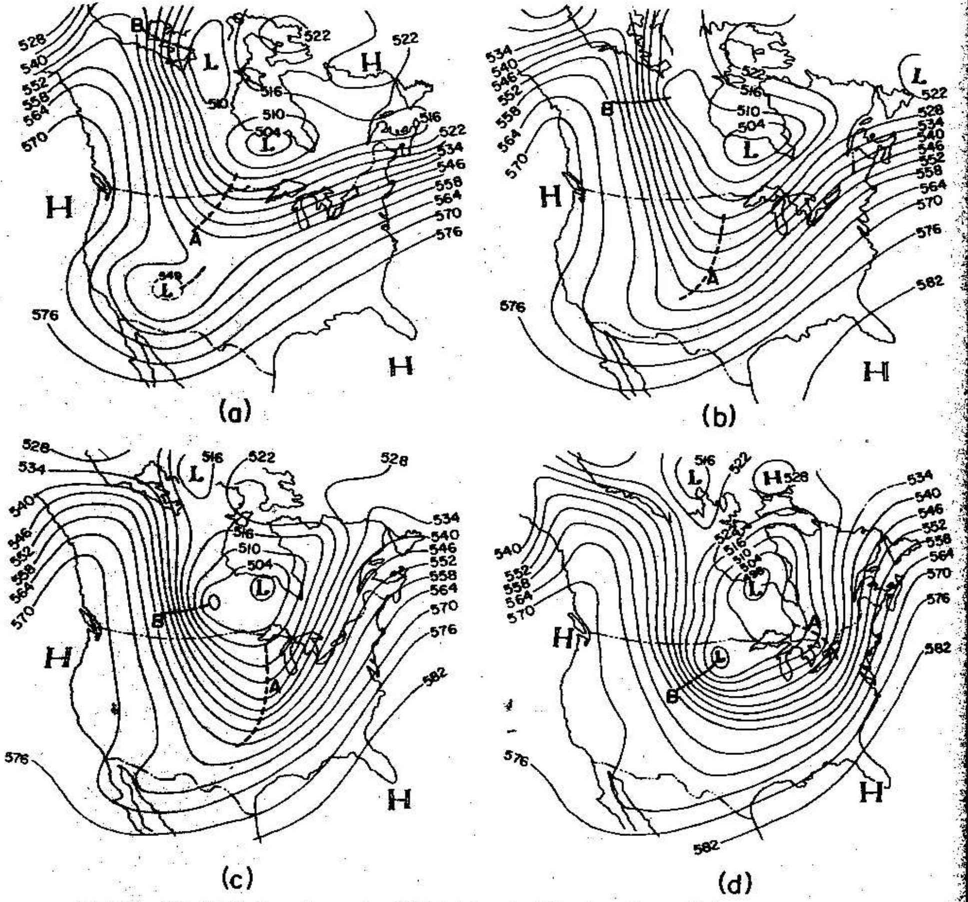


Fig. 3.2 The distribution of geopotential height on the 500-mb surface at 12-h intervals beginning at 00 GCT 19 November 1964 in (a) and ending at 12 GCT 20 November 1964 in (d). Contours are labeled in tens of meters.

are shown in Fig. 3.2. The feature of primary interest is the trough line labeled A in the figure. In chart (a) this feature is in the process of merging with a "cutoff low" that had been situated over the southwestern United States for several days. In charts (b) and (c) the combined trough line is swept eastward toward the Great Lakes. In chart (d), the trough line has come apart again; the northern segment has intensified and moved rapidly northeastward around the low center over central Canada, and the southern segment has been left behind. A second trough line, labeled B in the figure, can be tracked as it moves southward from the Canadian Arctic in (a) into the Northern Plains of the United States in (d). Considerable intensification of this feature can be noted during the 36-h period.

We will examine the upper level structure of this system in more detail in Section 3.4. First, however, we wish to describe the surface weather that occurred in association with trough line A as it moved eastward across the United States.

3.2 SURFACE WEATHER ELEMENTS

In the description of the surface weather, we will begin by confining our attention to the charts for 00 and 12 GCT 19 November and 00 GCT 20 November 1964, the counterparts of (a), (b), and (c) in Fig. 3.2. We will discuss the weather from the standpoint of various observed parameters: wind and pressure, temperature, dew point, precipitation, and pressure tendency. Then, having described the behavior of the various parameters separately, we will show how they are plotted together on a surface synoptic chart, or in the form of time sections at individual stations. In the course of this discussion we will introduce the concept of *fronts*, which is fundamental to the understanding of middle-latitude weather.

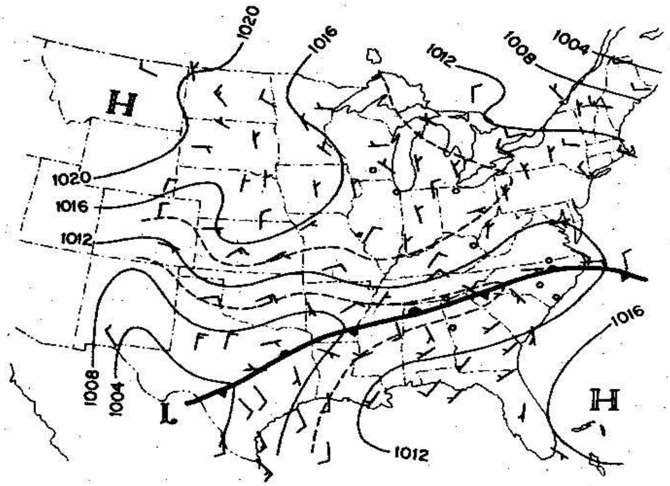


Fig. 3.3a Sea level pressure and surface winds at 00 GCT 19 November 1964. Heavy lines with barbs and half circles denote confluence lines. Lighter lines represent isobars drawn at 4-mb intervals for solid lines, 2-mb for dashed lines. The letters H and L denote centers of high and low pressure, respectively. Winds are plotted using the conventions described in Table 3.1. Small circles denote calm winds.

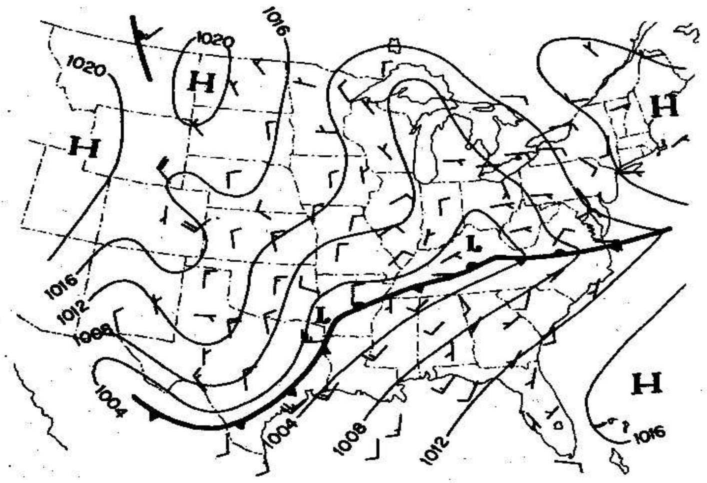


Fig. 3.3b Sea level pressure and surface winds at 12 GCT 19 November 1964. For further details, see caption of Fig. 3.3a.

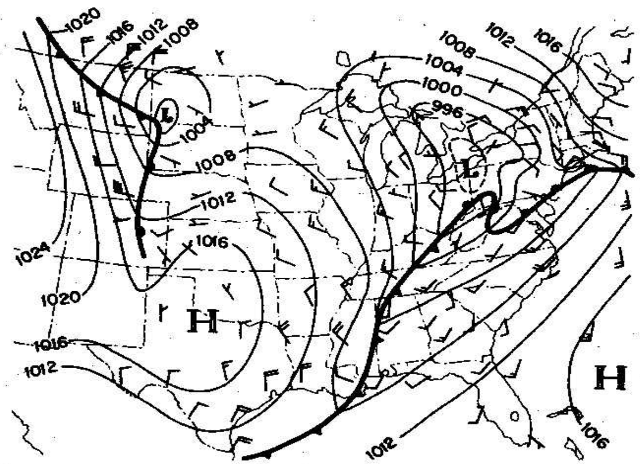


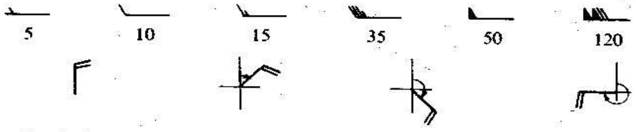
Fig. 3.3c Sea level pressure and surface winds at 00 GCT 20 November 1964. For further details, see caption of Fig. 3.3a.

3.2.1 Wind and pressure

In Fig. 3.3 the surface winds at individual stations are plotted vectorially, using the conventions described in Table 3.1. On the same charts are shown analyses of the sea level pressure, based upon data taken at the same stations as the wind observations. From a cursory inspection of Fig. 3.3 it is apparent

Table 3.1

Abbreviated plotting code for synoptic charts



Wind speed[a]	5	10	15	35	50	120

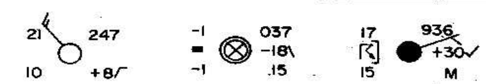
Wind direction	Northerly (from the north) (0° or 360°)	Northeasterly (45°)	Southeasterly (135°)	Westerly (270°) (southern hemisphere: note orientation of barbs)

Sky cover	○	⦶	◑	●	⊗
	Clear	Scattered clouds	Partly cloudy	Cloudy	Sky obscured (outer circle denotes calm wind)

Weather	Light continuous rain	Moderate continuous snow	Rain shower	Past drizzle	Thunderstorm	Dense fog	Sleet or hail	Freezing drizzle

Plotting model

Examples



ff	Wind speed (kt)	15	Calm	5
dd	Wind direction	320	—	070
TT	Temperature (°C)	21	−1	17
T_dT_d	Dew point (°C)	10	−1	15
PPP	Pressure	1024.7	1003.7	993.6
±ppa	Pressure tendency	0.8↑ (rising, then steady)	1.8↓	3.0↑ (fall, then larger rise)
N	Sky cover	Clear	Obscured	Cloudy
WW	Weather	None	Dense fog	Thunderstorm in past hour
RR	6 h precipitation (in.)	None	0.15	Missing

[a] Units: knots (1 knot = 1 nautical mile per hour; 1.95 knots = 1 m s^{-1}). 60 nautical miles = 1 deg of latitude; therefore, knots can easily be related to displacement measured in degrees of latitude over a 12- or 24-h period (for example, 10 knots = 4 deg day^{-1}).

that the surface wind tends to blow parallel to the isobars of sea level pressure, *l*eaving *l*ow pressure to the *l*eft, in accordance with the geostrophic relationship, but there is also some tendency for flow across the isobars from higher toward lower pressure because of the effects of friction.

On the first of the three charts, centers of high pressure are located over the Northern Plains and over the Atlantic Ocean east of Florida. Separating these regions of high pressure is an elongated pressure minimum or "trough" extending from the low center on the Texas–Mexico border, eastward across the Gulf States to North Carolina. This trough in the pressure field coincides with a line of *confluence* (flowing together) in the wind field, which separates a broad southwesterly air current streaming out of the Gulf of Mexico from a northeasterly flow streaming clockwise around the high pressure area over the northern plains. This confluence line is indicated by the heavy line with the pointed and rounded symbols. (The meaning of the symbols will be explained presently.)

On chart (b) in Fig. 3.3, which represents conditions 12 h later, wavelike undulations have developed along the confluence line. The crest of the first "wave" is located over eastern Kentucky, adjacent to the weak pressure minimum designated by the L, and the crest of the second wave is located over southwestern Arkansas, also adjacent to a weak minimum in the pressure field. The northerly flow to the west of the second wave crest appears to be sweeping the confluence line southward through Texas. Further to the east there has been a general northward shift of the confluence line over the 12-h period. The trough along the confluence line has deepened slightly, particularly in the vicinity of the wave crests.

By the time of map (c) in Fig. 3.3, more substantial changes have taken place. The first of the two waves on the confluence line has amplified to become the dominant one. Its crest has moved northeastward into southwestern Pennsylvania, and the associated center of low pressure has deepened by about 10 mb over the 12-h period. A distinct counterclockwise (cyclonic) circulation has developed around the deepening low center. To the east of the wave crest the confluence line is continuing to drift northward, allowing the Gulf air to advance into the Middle Atlantic States, while to the west of the wave crest the cyclonic circulation is sweeping the confluence line southeastward, in advance of the flow of air from the Great Plains. A second confluence line has moved southeastward out of western Canada into the northern plains of the United States. This line is most distinct to the west of the low center, where it separates a northerly flow of Arctic air from a westerly flow of air from the Pacific. The deepening low center over southwestern North Dakota is located near the crest of a wave on this second confluence line.

3.2.2 Temperature; fronts

The distribution of temperature at 00 GCT 19 and 20 November 1964 (which correspond roughly to 6 p.m. local time, 18 and 19 November 1964, respectively) is shown in Fig. 3.4. Here we have chosen to show the raw temperature

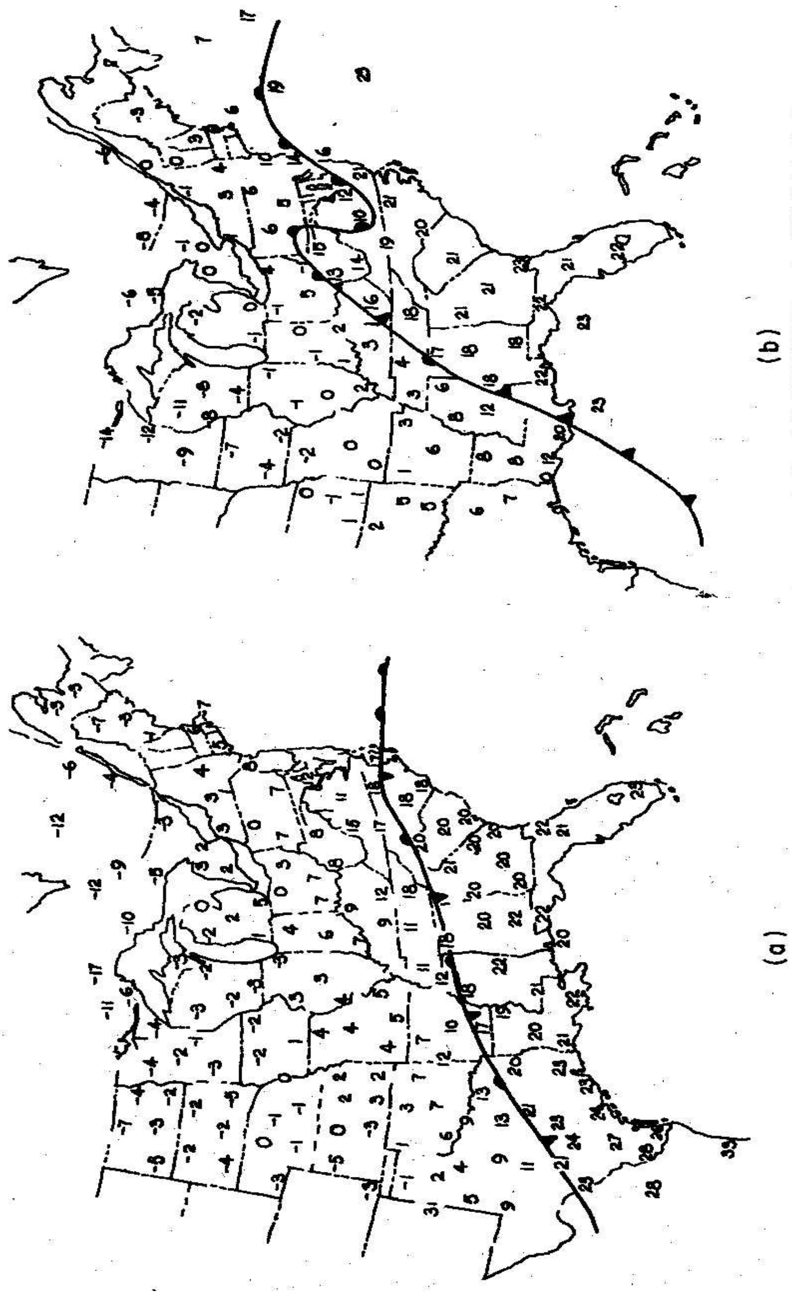


Fig. 3.4 Surface temperatures in degrees Celsius and frontal positions: (a) 00 GCT 19 November 1964, and (b) 00 GCT 20 November 1964. Pointed barbs denote cold fronts, half circles denote warm fronts, and alternating pointed barbs and half circles denote stationary fronts.

data, rather than an analysis of the distribution of surface isotherms. The confluence lines from Fig. 3.3a and c have been transcribed onto these charts.

In Fig. 3.4a, the temperatures to the south of the confluence line are uniformly quite high for the season, which is understandable in view of the fact that the air over this region formerly resided over the warm surface waters of the Gulf of Mexico. Within this warm air mass there is relatively little horizontal temperature gradient; the temperatures along the confluence line are almost as high as they are along the Gulf Coast.

There is no discontinuity in the value of the temperature as one crosses the confluence line. However, within the first 100–200 km to the north of the line there is a band of very strong, horizontal temperature contrast, with temperatures dropping from values of about 20°C along the confluence line itself to about 10°C some 100–200 km to the north of the line. Proceeding further toward the north, the temperatures continue to drop, but at a slower rate. Thus, the confluence line defines the boundary between a rather homogeneous warm air mass and a region of strong thermal contrast between the warm air mass and the colder air to the north. Because of its role as a boundary or dividing line, the confluence line is called a *front*. The region of strong thermal contrast on the "cold air side" of the confluence line is called a *frontal zone* (or sometimes, a *baroclinic zone*).

It should be emphasized that in the context of this definition, it is not quite correct to say that a front marks the boundary between a warm air mass and a colder air mass. The transition between the two air masses takes place within a zone of finite width—the frontal zone. The front is the warm air boundary of the frontal zone and it coincides with the line of confluence in the wind field.

Fronts are labeled in terms of their direction of movement. If the air on the cold side of the frontal zone is advancing into the region formerly occupied by warmer air, the front is called a *cold front*. Cold fronts are denoted on weather charts by triangular shaped "teeth" which point in the direction of movement. For example, the front over the southeastern United States in Figs. 3.3c and 3.4b is moving southeastward as a cold front. Similarly, if the air on the cold side of the front is retreating and being replaced by warmer air, the front is called a *warm front* and denoted on charts by semicircular symbols which point in the direction of frontal movement; in this case, in the direction of the colder air. For example, note the front along the eastern seaboard in Figs. 3.3c and 3.4b. *Stationary fronts* are denoted by alternating cold and warm front symbols on different sides of the line.

To a rather close approximation, fronts behave as *material surfaces* in the atmosphere; that is to say, if one could somehow tag or label the air parcels that lie along a frontal surface at some instant in time and follow them as they move along their respective three-dimensional trajectories through space, these very same air parcels would continue to define the frontal surface at future times. Thus it is almost correct to say air does not move through a frontal surface.

In order to understand why fronts move as they do, it is useful at this point to consider briefly their vertical structure, as depicted schematically in Fig. 3.5, which shows vertical cross sections normal to fronts which are exhibiting various types of movement. All three sections are drawn such that the warm air lies toward the left and the colder air toward the right. Note that regardless of the direction of movement the frontal surface slopes in the direction of the cold air with increasing height. Thus, in all three cases, the frontal zone lies below the frontal surface and the warm air mass lies above it.[†] Thus it is possible for the warm air to be lifted up and over the frontal surface as shown in Fig. 3.5. In contrast, the air within the frontal zone is "trapped" in the shallow wedge beneath the frontal surface, and thus cannot move relative to the front, or, conversely, the front cannot move relative to it. Hence the direction and speed of movement of the front is determined by the winds within the frontal zone. For example, in Fig. 3.3(c) the front over the southeastern United States is moving southeastward as a cold front, pushed forward by the northwesterly winds within the frontal zone, while the warm front over the eastern seaboard is moving northward, following the retreat of the frontal zone air in that region. It is extremely important to be aware of the need for consistency between frontal movement and the wind field when analyzing sequences of synoptic charts.

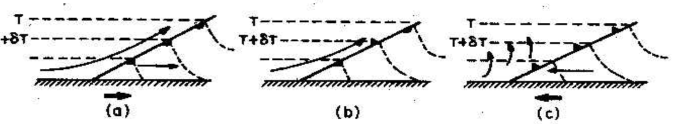


Fig. 3.5 Idealized vertical cross section through frontal zones showing isotherms (----) and air motions relative to the ground (→). (a) Warm front, (b) stationary front with overrunning warm air, and (c) cold front. Heavy arrows at bottom indicate sense of frontal movements.

Lest we overemphasize the role of fronts as a factor in the variability of surface temperature, it should perhaps be mentioned that other factors such as time of day, sky cover, altitude of the station, and proximity to the ocean can be equally, or more, important at times. In fact, there are large regions of the globe in which it is extremely difficult to locate fronts on the basis of gradients of surface temperature:

- over the oceans, where the surface temperature never departs by more than a few degrees from the temperature of the underlying water,
- in mountainous terrain where large differences in station elevation introduce spurious temperature gradients.

[†] In the case of rapidly moving cold fronts the lowest portion of the front is sometimes retarded by friction so that the frontal zone "overhangs" a narrow strip of warm air at the earth's surface. This effect is usually restricted to the lowest few hundred meters of the atmosphere.

In the analysis of surface charts it is essential to discriminate between the temperature gradients associated with fronts and those due to other influences.

3.2.3 Dew point

Just as frontal zones are characterized by strong horizontal temperature gradients, they also tend to be marked by strong horizontal gradients of dew point, especially when the cold air is of continental origin and the warmer air is of maritime origin, as is often the case over the central and eastern United States. In the 19–20 November 1964 case the distributions of temperature and dew point are so similar that we will not take the space to show the dew points in relation to the front. However, in certain synoptic situations, the dew point gradient is much more reliable than the temperature gradient as an indicator of frontal positions. For example, during summer over land the diurnal temperature range at the ground tends to be considerably larger in cool, dry continental air than in warm, moist air off the Gulf of Mexico, which is often characterized by cloudy skies. Thus, during afternoons, it is not uncommon for surface temperatures in the "cold" air to be just as high as those on the "warm" side of the front, even though there is still considerable thermal contrast 1 km above the ground. Under such conditions the horizontal gradient of dew point is likely to be quite large in the frontal zone, thus providing a clear indication of the frontal position. If such a frontal zone passes a station, moving from northwest to southeast, the observer is likely to notice that the air is becoming much less humid, even though the temperature remains fairly high.

3.2.4 Precipitation

The distribution of precipitation and fog is closely related to the frontal positions as shown in Fig. 3.6. In Fig. 3.6a light rain and snow are falling throughout a broad band to the north of the stationary front as warm air overruns the sloping frontal zone, as indicated in Fig. 3.5b. The light snow extends farther northward along the slopes of the Rockies because of the easterly low-level flow which induces additional lifting in this region where the terrain slopes upward toward the west.

In the 12-h period between (a) and (b) in Fig. 3.6 the precipitation has increased both in intensity and in areal coverage. Snow, sleet, and rain have spread northward in response to increased overrunning above the warm front. Meanwhile an extensive area of fog has developed in the vicinity of the warm front itself. Fog is common wherever warm, moist air passes over a colder, underlying surface. The band of precipitation associated with the cold front is narrow, but rather intense in places, as evidenced by the locally heavy precipitation amounts. Many of the stations in the vicinity of the front reported thunderstorms during the previous 12-h period. In Section 5.5 we will consider in more detail the distribution of precipitation in extratropical cyclonic storms.

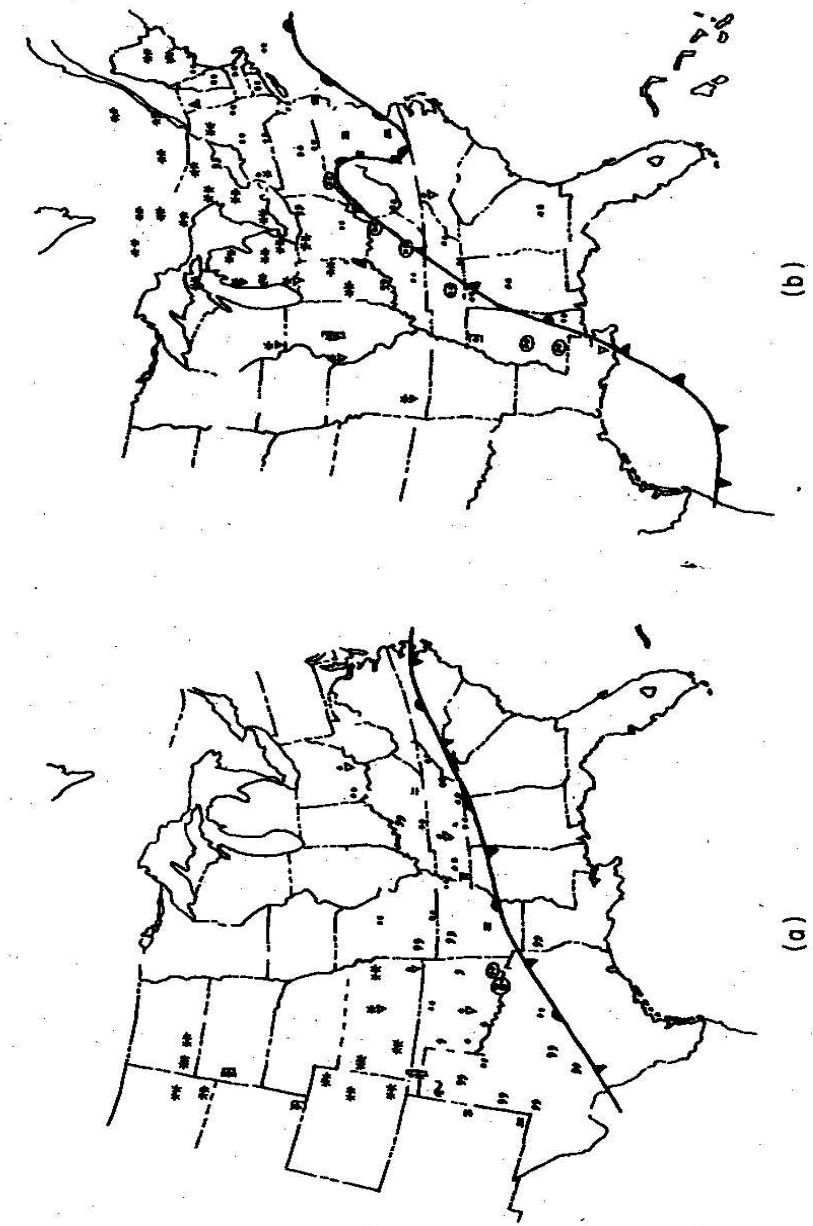


Fig. 3.6 Distribution of precipitation in relation to fronts: (a) 00 GCT 19 November 1964, and (b) 00 GCT 20 November 1964. For an explanation of plotting conventions, see Table 3.1. Symbols with circles around them denote stations that have recorded more than 1 cm of precipitation during the past 6 h.

3.2.5 Pressure tendency

The distribution of pressure change during the past 3 h provides an indication of the direction of motion and the rate of intensification or weakening of various features on the surface chart. The pressure tendency is particularly valuable as an aid in locating fast-moving fronts. For example, as a warm front approaches a station, the layer of relatively dense frontal-zone air adjacent to the ground gradually becomes thinner and warmer. Both effects contribute to a decrease in (hydrostatic) pressure at the ground. The passage of the warm front itself is usually accompanied by a leveling off of the pressure. In a similar manner, the passage of a cold front usually marks the beginning of a period of pronounced pressure rises at the ground. Cold fronts usually slope more steeply, relative to the ground, than warm fronts, and therefore the pressure rises that follow cold fronts tend to be more abrupt than pressure falls that precede warm fronts.

Figure 3.7 shows the distribution of surface pressure tendency at 00 GCT 20 November 1964, the time of the second map in the preceding figures. The lines connecting points at which the same pressure tendency occurs are called *isallobars*. Note the broad area of pronounced pressure falls in advance of the warm front and the narrow band of very strong pressure rises immediately behind the cold front.

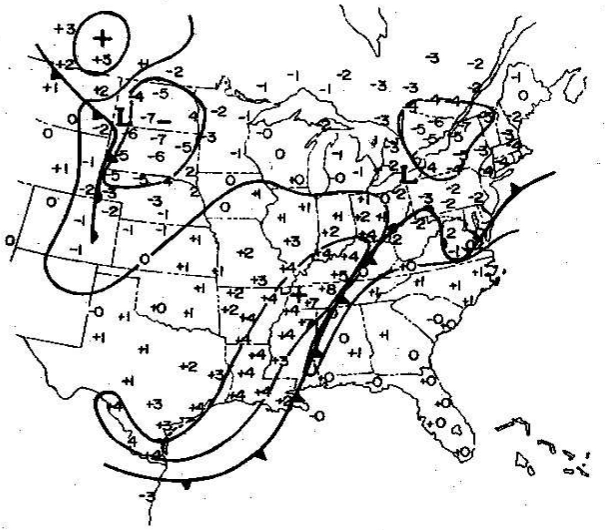


Fig. 3.7 Distribution of pressure change (in whole millibars) during the 3-h period ending at 00 GCT 20 November 1964. Isallobars are drawn at intervals of 4 mb $(3\ h)^{-1}$.

The distribution of pressure tendency also provides an indication of how the intensities of surface cyclones and anticyclones are changing with time. For example, in Fig. 3.7 the pressure is falling in the vicinity of the center of the low pressure area near Buffalo. Thus it is clear that the low is deepening as it moves northeastward and that the associated cyclonic circulation in the wind field is intensifying. In a similar manner the predominance of pressure rises near the center of a high indicates "building up" of the high and an intensification of the related anticyclonic circulation in the wind field.

When interpreting small changes in pressure, it should be borne in mind that the diurnal cycle in solar heating produces small but noticeable pressure fluctuations that have little or nothing to do with the synoptic situation. These "tidal" fluctuations should be subtracted out of the pressure tendencies before trying to infer rates of change that relate to the synoptic scale patterns. Since the tidal pressure changes are geographically and seasonally dependent, these corrections must be made on the basis of climatological data.

3.3 INTERPRETATION OF SYNOPTIC SURFACE REPORTS

An abbreviated version of the plotting model used for surface synoptic observations is shown in Table 3.1. This model incorporates all the meteorological parameters already discussed. The full surface synoptic report contains additional information on visibility, dominant cloud types, and height of the base of the lowest cloud layer, which we will not discuss here. There is also a more extensive set of symbols for describing weather in more detail and for reporting other types of restrictions to visibility such as haze, smoke, dust, and blowing snow. Most of this additional information carried in the surface reports is included primarily because of its applications to aviation. Ship reports include observations of sea surface temperature (or sea–air temperature difference) and a specification of the present rate of movement of the ship.

3.3.1 The synoptic surface chart

The conventional surface chart contains an analysis of the sea level pressure field together with the positions of fronts. In positioning fronts and centers of high and low pressure on the surface chart, the analyst strives to integrate the information plotted for the individual meteorological parameters in order to obtain a representation that is fully consistent with the current synoptic data, and with the previous synoptic chart. A certain amount of subjectivity is inherent in this process, but fortunately there is usually enough redundancy between successive charts, and between the fields of different meteorological parameters, to eliminate most of the ambiguities. In regions of sparse surface synoptic coverage, satellite imagery is the primary basis for positioning the major features on the surface chart.

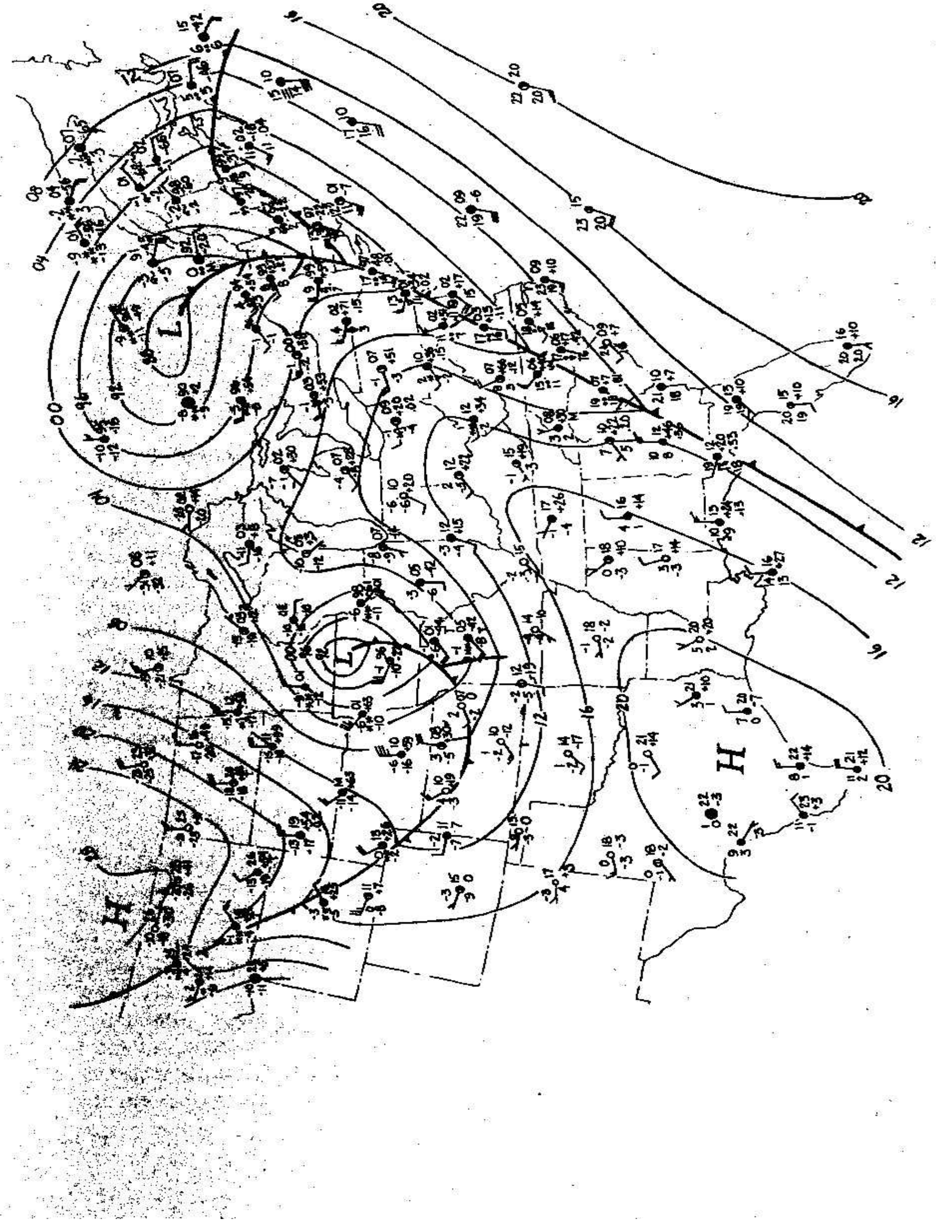


Fig. 3.8 Surface synoptic chart at 12 GCT 20 November 1964. Plotting conventions are as outlined in Table 3.1.

As an example of a surface synoptic chart we show in Fig. 3.8 the analysis for 12 GCT 20 November 1964. Certain familiar features can be identified in this later chart. The area of low pressure that was over the eastern Great Lakes at 00 GCT has moved northeastward into Quebec and deepened slightly, while the wave along the warm front that was located south of Rhode Island has moved northeastward to the coast of New Brunswick. The cold front has progressed rapidly eastward during the 12-h period, and is now approaching the Atlantic coastline.

With the passage of time, the center of lowest pressure is becoming more and more detached from the region of the frontal zone, where it originally formed, as it propagates northward into the cold air. In this latest chart there is evidence of a line of discontinuity connecting the low center over Quebec with the junction of the warm and cold fronts in Connecticut. Stations located ahead (to the east) of the line are experiencing typical pre-warm-front weather (rising temperatures, easterly winds, and falling pressures), while stations located behind (to the west of) the line are experiencing typical post-cold-frontal weather (falling temperatures, decreasing precipitation, gusty westerly winds, and sharply rising pressures). Thus at most stations the passage of this line is marked by a temperature maximum, a pressure minimum, and a windshift. The intensity of the "back to back" frontal zones on either side of the line of discontinuity is directly related to the warmth of the air along the line itself. The thermal contrast is strongest near the junction of the warm and cold fronts, where the air along the line of discontinuity is quite warm, and it becomes progressively weaker as one moves northward along the line, toward the colder air at the center of the low. It is customary to indicate such lines of discontinuity by alternating cold and warm front symbols pointing in the direction of motion, and to refer to them as *occluded fronts*, *occlusions*, or *trowals* (that is, *trough*, *warm air aloft*; a Canadian term).

3.3.2 Time series representation

Figure 3.10 shows time series of surface reports at selected stations in eastern United States and Canada on 19–20 November 1964. (Locations of the stations are shown in Fig. 3.9.) The stations are arranged according to latitude, from north to south. The sequences may be summarized as follows:

- The southernmost stations BR (Brownsville, Texas), LC (Lake Charles, Louisiana), and JA (Jackson, Mississippi) are in the warm air at the beginning of the period. There is a distinct cold front passage accompanied by a distinct windshift and followed by a period of falling temperatures and dew points, and sharply rising pressures. The frontal passage is without precipitation at Brownsville, but it is accompanied by showers and thundershowers at Lake Charles and Jackson.
- NA (Nashville, Tennessee) and HT (a composite of Huntington and Charleston, West Virginia), the next stations to the north, were within the frontal

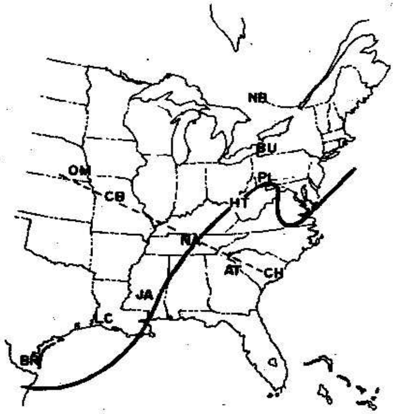


Fig. 3.9 Location of stations used in time sections (Fig. 3.10), soundings (Fig. 3.18), and vertical cross section (Figs. 3.19 and 3.20): (---) orientation of the cross section; (——) the frontal positions at 00 GCT 20 November 1964, the time of the soundings and cross section. Station names are given in the text.

zone at the beginning of the sequence. They experience a period of rising temperatures and falling pressures during the time that the front drifts northward toward them. Then there follows a brief period of southwesterly winds and high temperatures, which begins with the passage of the warm front shortly after 18 GCT 19 November. (The interval of time within the warm sector of the storm is longer for stations located farther to the east.) The brief warm interval is followed by a sharp cold front passage.

- The warm air never reaches PI (Pittsburgh, Pennsylvania) and yet there is evidence of a frontal passage shortly after midnight GCT 20 November. Prior to this time, conditions are similar to those at HT and NA before the warm front passage, and afterwards the sequence is similar to those at HT and NA following the passage of the cold front. Therefore, the front that passed PI has characteristics typical of an occlusion, as described above. Indeed, an occluded front could have been drawn just west of Pittsburgh in Fig. 3.3c.
- The sequence of events at BU (Buffalo, New York) is not as clearly defined as at PI. There is a windshift and a pressure minimum around 03 GCT 20 November, but there is no well-defined temperature maximum at this time. However, prior to the windshift there is evidence of warming aloft, with snow changing to rain, while some time after the windshift the rain changes back to snow. Conditions at BU are indicative of the passage of a weak occluded front, far from the junction of the warm and cold fronts.
- NB (North Bay, Ontario) is situated deep within the cold air throughout the time sequence. There is a pressure minimum and a windshift as the center

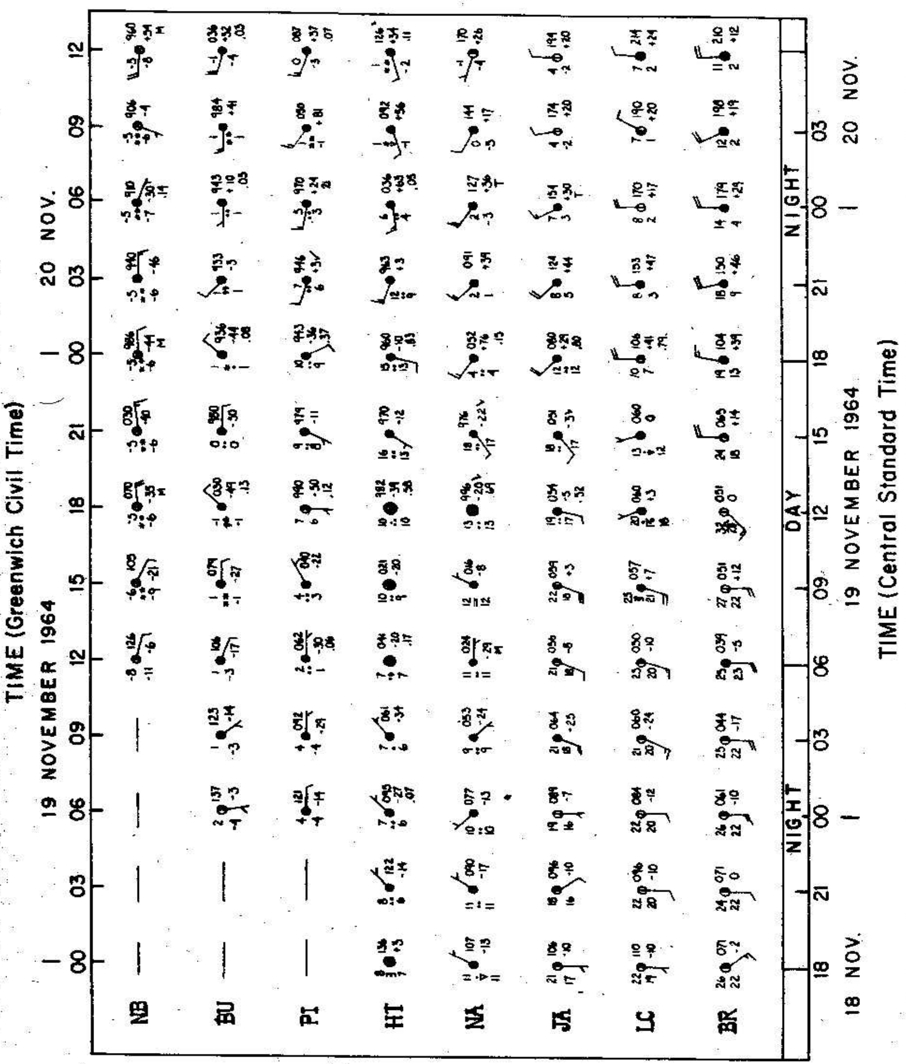


Fig. 3.10 Time series of surface synoptic reports. Station locations are shown in Fig. 3.9, plotting conventions in Table 3.1.

of the low passes the station, but these changes are not associated with any frontal passage.

3.3.3 Models of the life cycle of extratropical cyclones

Most middle-latitude cyclones have much in common with the "classical textbook example" described in this chapter. Important characteristics of such "polar front cyclones" are

- the initial development of the low pressure center along a stationary front; the low center develops on the crest of a wavelike undulation in the shape of the front;
- the ensuing movement of the frontal zone in response to the developing circulation around the deepening low pressure center, with the cold air retreating toward higher latitudes in advance of the surface low, and sweeping equatorward and eastward behind it;
- the propagation of the low pressure center toward the cold air as it deepens, with an occluded front connecting the low center to the junction of the warm and cold fronts; the occlusion process usually marks the end of the period of rapid development (cyclogenesis).

The sequence of events described above is embodied in the idealized model shown in Fig. 3.11, which is very similar to the one first developed by the Bergen school[†] more than 50 yr ago. This model has been widely used by weather forecasters as a basis for interpreting and anticipating changes in the surface synoptic chart.

[†] The school was founded in 1918 by the Norwegian physicist **Vilhelm Bjerknes** (1862–1951), his son **Jacob Bjerknes** (1897–1975), **Halvor Solberg** (1895–), and **Tor Bergeron** (1891–). The elder Bjerknes began his career as a physicist. In the early 1890s he collaborated with Heinrich Hertz and published several fundamental papers in radio science. In the latter part of that decade he turned his attention to the dynamics of atmospheres and oceans. The "circulation theorems," which he developed during this period, provide a theoretical basis for the basic concepts discussed in Section 9.3 of this book. During World War I, when Norway was cut off from most outside weather information, Bjerknes was called upon to found a Geophysical Institute at Bergen. In this role he was successful in convincing the Norwegian government to install a dense network of surface stations which provided data for investigating confluence lines in the surface wind field. These studies led to the concept of fronts and ultimately to models of the life cycle of frontal cyclones. In his characteristically modest manner, Bjerknes credited his younger colleagues with the major scientific achievements of the Bergen school: "During 50 years meteorologists all over the world had looked at weather maps without discovering their most important features. I only gave the right kind of maps to the right young men, and they soon discovered the wrinkles in the face of Weather."

In 1919, J. Bjerknes (aged 22 at the time) published an eight-page paper which introduced the concept of warm, cold, and occluded fronts and correctly explained their relationship to extratropical cyclones. By 1926, in collaboration with Solberg and others, he had described the structure and life cycle of extratropical cyclones. Bergeron made important contributions to the understanding of occluded fronts and the formation of precipitation (see Section 4.5.4).

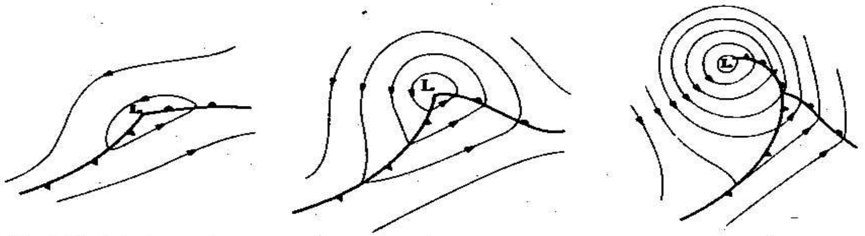


Fig. 3.11 Idealized model of a middle-latitude cyclone in three stages of development showing isobars of sea level pressure and fronts. Arrows indicate the direction of the geostrophic wind. (Adapted from E. Palmén and C. W. Newton, "Atmospheric Circulation Systems," Academic Press, New York, 1969.)

It should be emphasized that relatively few extratropical storms fit the idealized "polar front cyclone" model as well as the one selected for display in this chapter. Clearly defined warm fronts are lacking in many storms,[†] particularly during the later stages of development, and topographical features often distort or obscure existing fronts,[‡] or generate new fronts that would not exist otherwise. Furthermore, there is increasing evidence that some storms develop in isolation from any pre-existing fronts; the fronts that accompany these systems form as part of the process of storm development. Some of these forms of "aberrant behavior" are difficult to understand or predict on the basis of the surface synoptic map alone, but they make sense when viewed as a response to changes that are taking place at upper levels.

3.3.4 Further remarks on occluded fronts

Occluded fronts have been subject to various conflicting interpretations. The term *occluded* means overlapping: it stems from the widely held notion that such fronts form when the cold front "catches up with" part of the warm front during the process of cyclogenesis. When actual frontal movements are carefully examined, there are few, if any, well-documented examples of cold fronts overtaking warm fronts to form occlusions. Rather, it appears that most occluded fronts are essentially new fronts which form as surface lows separate themselves from the junctions of their respective warm and cold fronts and deepen progressively further back into the cold air.

The concept of an occluded front can be understood in terms of the idealized model shown in Fig. 3.12, which stems from the Bergen school. In the cold-type

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[†] Note, for example, the storm over the Northern Plains in Fig. 3.8.

[‡] The southward bulge of the warm front in the Middle Atlantic States in Fig. 3.3c is an effect of the Appalachian Mountains. Most warm fronts display a similar distortion when they pass over this region.

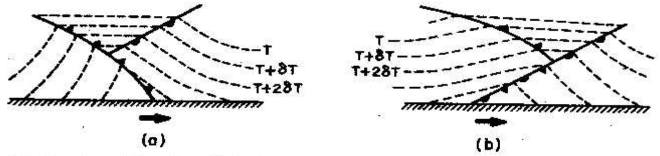

Fig. 3.12 Idealized models of occluded fronts: (a) cold type and (b) warm type. The sketches represent frontal surfaces (——) and isotherms (---) in vertical cross sections normal to occluded fronts, which are moving from left to right.

occlusion shown in Fig. 3.12a the cold front extends to the ground and the warm front exists only aloft, whereas in the warm-type occlusion (Fig. 3.12b) the reverse is true. Note that both types of occlusion are consistent with the notion of "back to back" frontal zones at the ground, so that weather conditions prior to the frontal passage are similar to those ahead of a warm front and conditions after the frontal passage are similar to those behind a cold front. It is clear from the diagrams that the passage of a "cold-type occlusion" is marked by an increase in the static stability of the lower troposphere, whereas a warm-type occlusion is marked by a decrease in static stability.

The classical occluded structures shown in Fig. 3.12 are rarely observed in their entirety. One or more of the frontal discontinuities indicated in the figure is often missing or obscured by mesoscale features in the vicinity of the front (as will be discussed in Section 5.5). For example, there may not be a well-defined frontal passage at the ground, or the warm (or cold) frontal zone aloft may lack a well-defined warm air boundary. In such situations it may not be clear whether the occlusion is of the warm or cold type. In view of the wide variety of occluded frontal structures that exist in nature, it is advisable to identify occluded fronts, not in terms of a set of models of frontal configurations but rather in terms of their essential characteristics: namely,

- back to back frontal zones at low levels with the warmest air in the vicinity of the front,
- a trough in sea level pressure.

It will be shown in Section 3.5.1 that an occluded front can also be identified in terms of a ridge in the lower tropospheric thickness field.

3.4 UPPER LEVEL STRUCTURE

The time evolution of the patterns on the surface synoptic chart becomes more understandable when viewed in the context of the synoptic situation in the troposphere as a whole. For example, it is observed that the winds in the middle troposphere (near 500 mb) tend to act as a "steering flow" for features on the surface chart. The upper level patterns also influence the rate of intensification or weakening of surface cyclones and anticyclones, and the amount and type of precipitation that accompanies them.

3.4.1 Upper level synoptic charts

Figures 3.13–3.17 show the distributions of geopotential height and temperature on the 850-, 700-, 500-, 250-, and 100-mb pressure surfaces, respectively, at 00 GCT 20 November 1964 (the same time as the surface charts shown in Figs. 3.3c, 3.4b, and so on).

The 850-mb chart is rather similar to the surface chart discussed previously, but there are some notable differences:

- The closed lows over Buffalo, New York, and western North Dakota at the surface appear as troughs at the 850-mb level.

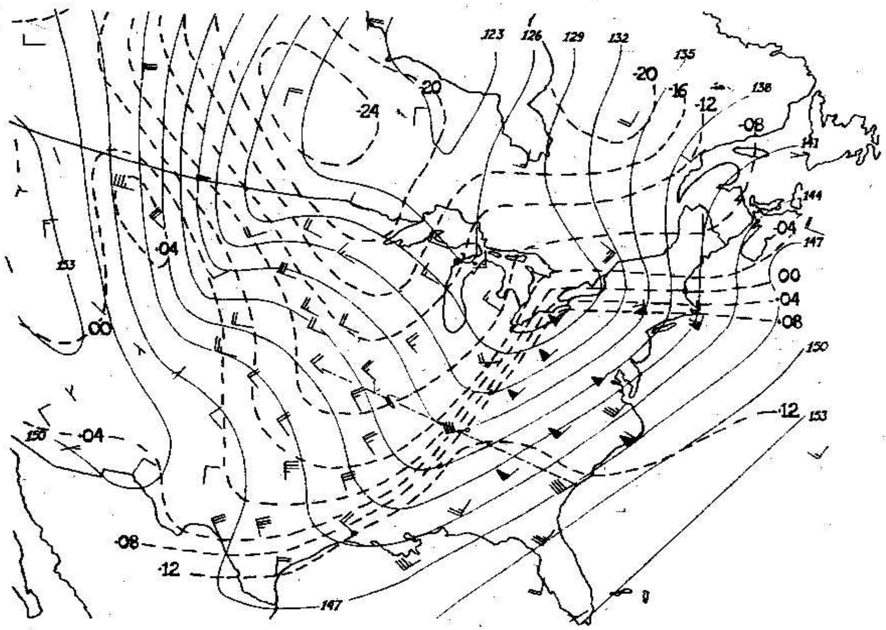

Fig. 3.13 850-mb chart for 00 GCT 20 November 1964: (——) geopotential height contours, drawn at intervals of 30 m and labeled in tens of meters; (---) isotherms, labeled in degrees Celsius. Wind speeds are in knots. The stations denoted by the small circles are Lake Charles and Nashville.

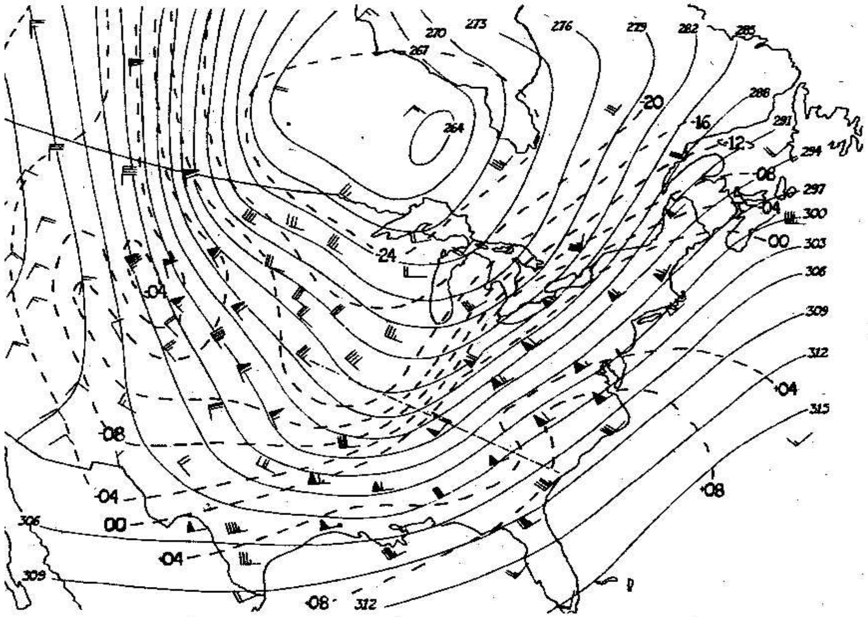

Fig. 3.14 700-mb chart for 00 GCT 20 November 1964: (——) geopotential height contours, drawn at intervals of 30 m and labeled in tens of meters; (– – –) isotherms, labeled in degrees Celsius. Wind speeds are in knots. The stations denoted by the small circles are Lake Charles and Nashville.

- The winds are generally stronger at 850 mb than at the ground, and there is no evidence of systematic flow across the isobars toward lower pressure as there is on the surface map.
- The cold front (defined as the warm air boundary of the frontal zone) is well past NA (Nashville) and LC (Lake Charles) on the surface chart, but it has just reached them at 850 mb. Thus, the cold front slopes backward toward the cold air with increasing height, in agreement with the model described in Fig. 3.5.
- The warm front is more clearly defined and much farther north at 850 mb than on the surface map. Apparently the "pool" of cold (frontal zone) air over the Middle Atlantic States (see Fig. 3.4b) is very shallow.

Proceeding upward from 850 to 700 mb we note the following:

- The troughs associated with the surface low centers are becoming less distinct and they are displaced upstream relative to their positions on the surface map.

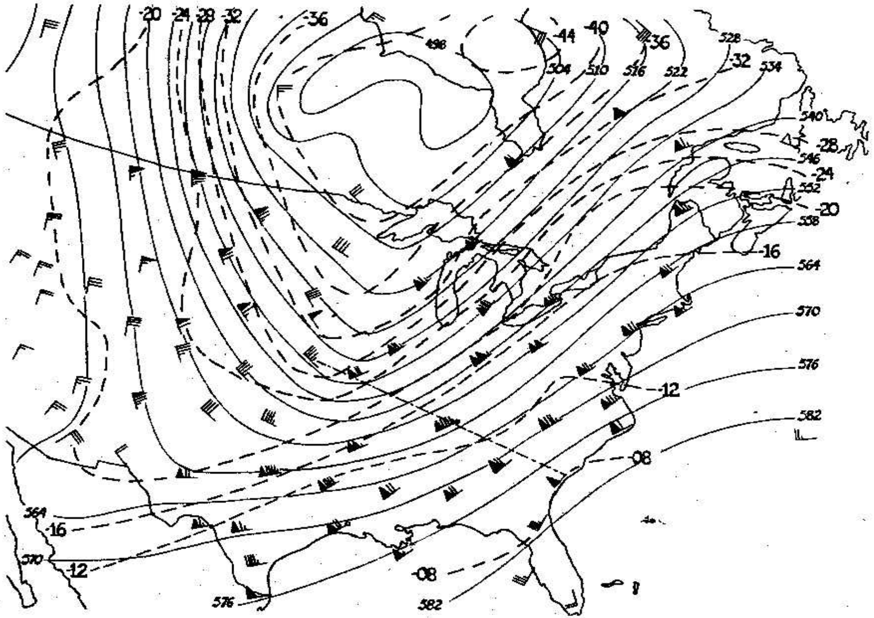

Fig. 3.15 500-mb chart for 00 GCT 20 November 1964: (——) geopotential height contours, drawn at intervals of 60 m and labeled in tens of meters; (– – –) isotherms, labeled in degrees Celsius. Wind speeds are in knots.

- There is a further increase in wind speed with height at most stations.
- The cold front is located still farther toward the northwest of its surface position (note that it has not yet reached NA and LC).
- The warm front is located far to the north of its 850-mb position. Much of the warm front precipitation in Fig. 3.6b coincides with the upper level position of the frontal zone.
- The frontal zones are less clearly defined than at lower levels. For example, it is difficult to identify the southern portion of the cold front at this level.

The transition from the 700-mb chart to the 500-mb chart is marked by a further upstream displacement of the troughs, strengthening of the winds, and sloping of the frontal zones toward the cold air. At the 500-mb level, well-defined frontal zones are rather unusual, but strong thermal contrasts still exist. Note that the gross features of the 500-mb temperature field are similar to those on the charts for lower levels. In fact, throughout the depth of the troposphere, the isotherms have much the same orientation.

We recall from Figs. 1.10 and 1.11 that, in a climatological sense, there is in middle latitudes a distinct break between the high, cold tropical tropopause

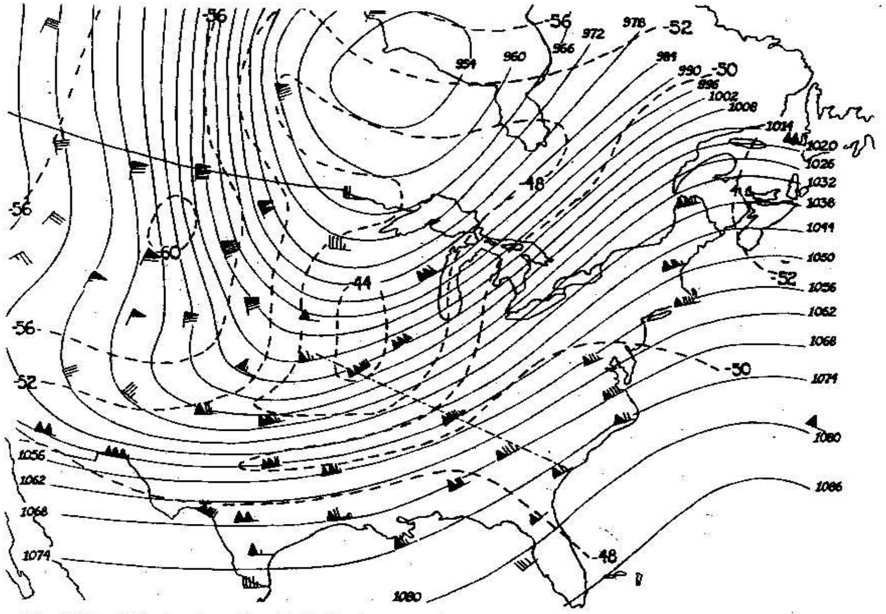

Fig. 3.16 250-mb chart for 00 GCT 20 November 1964: (——) geopotential height contours, drawn at intervals of 60 m and labeled in tens of meters; (---) isotherms, labeled in degrees Celsius. Wind speeds are in knots.

and the lower, warmer polar tropopause. It was shown in Fig. 1.11 that this break coincides with the climatological position of the jet stream. The 250-mb chart shown in Fig. 3.16 cuts across this tropopause break at the position of the jet stream. North of the jet stream the 250-mb level is located in the lower stratosphere, while south of the jet stream it is located in the upper troposphere. The warmest air at 250 mb is located on the poleward side of the jet stream in the region of the trough, over the central United States. It will be shown that this warm air coincides with a region of very low tropopause heights.

The 100-mb surface is located well above the tropopause and jet stream. In passing upward into the stratosphere there is a marked decrease in wind speeds and a change in the scale of the circulation patterns: the charts up to the 250-mb level were characterized by a superposition of synoptic-scale and planetary-scale features, whereas only planetary-scale features are present on the 100-mb chart. Note that the temperature field completely reverses between troposphere and stratosphere. For example, at the 100-mb level, the highest temperatures are found over central Canada and the lowest temperatures are found in the vicinity of Florida.

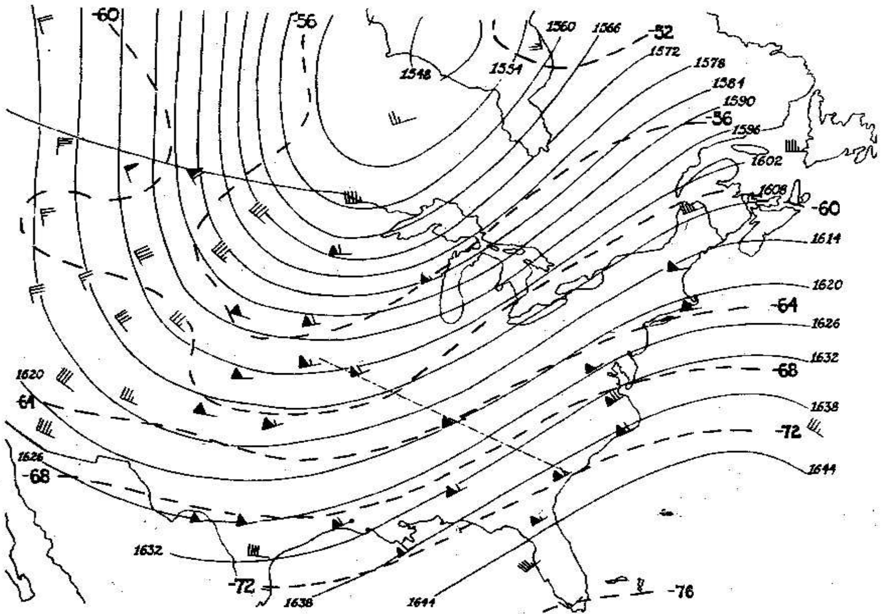

Fig. 3.17 100-mb chart for 00 GCT 20 November 1964: (——) geopotential height contours, drawn at intervals of 60 m and labeled in tens of meters; (---) isotherms, labeled in degrees Celsius. Wind speeds are in knots.

3.4.2 Vertical soundings

Vertical temperature soundings for four stations oriented along a line perpendicular to the cold front are shown in Fig. 3.18. The locations of the stations are shown in Fig. 3.9.

The sounding for Athens, Georgia (AT) shows the typical vertical structure of the warm air mass. With the exception of a few minor temperature inversions, the lapse rate is rather uniform and close to moist adiabatic throughout the troposphere. The tropopause is well defined and occurs at a level of about 180 mb or roughly 13 km. At lower latitudes the higher discontinuity in the temperature profile near 100 mb becomes the dominant tropopause.

The sounding for Nashville, Tennessee (NA) is rather similar to the one just described, except for the appearance of the stable frontal-zone air at low levels. The cold front coincides with the top of the inversion, at 850 mb. We recall from Fig. 3.13 that the cold front passes through NA at the 850-mb level, in agreement with the sounding. The increased static stability within the frontal zone is consistent with the thermal structure indicated in Fig. 3.5.

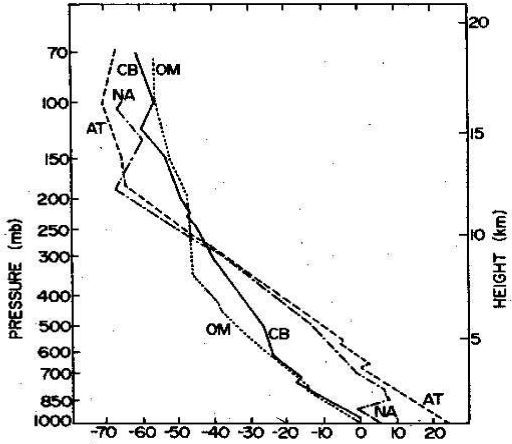

Fig. 3.18 Vertical temperature soundings at 00 GCT 20 November 1964. See Fig. 3.9 for the station locations.

Proceeding farther back into the cold air to Columbia, Missouri (CB) we begin to see more pronounced changes in the vertical temperature profile. The frontal zone is no longer clearly evident, but below the 300-mb level the whole sounding is much colder than the one for NA. The lapse rate gradually decreases from troposphere to stratosphere, but it is impossible to identify any distinct point on the profile at which the transition from troposphere to stratosphere takes place. Above 300 mb the air is warm relative to the first two soundings.

Omaha, Nebraska (OM) is located near the 500-mb trough deep within the cold air mass. On the whole, the sounding for this station is rather similar to the one for CB. The one notable difference is the reappearance of the tropopause, near 350 mb, or roughly 8.5 km. We note that in the climatological cross sections shown in Fig. 1.11 the tropopause does not reach as low as 350 mb, even over the poles. Such extremely low tropopause heights are found only locally in the vicinity of sharp upper tropospheric troughs. In general such regions tend to be characterized by low temperatures throughout the depth of the troposphere, but very high temperatures just above the tropopause. In general, there seems to be a tendency for compensation between tropospheric and lower stratospheric temperatures.

3.4.3 Vertical cross sections

The vertical structure of the atmosphere in the vicinity of the cold front is shown in Fig. 3.19. This cross section was constructed on the basis of temperature and wind soundings at the five stations indicated in the legend and constant pressure level charts such as those shown in Figs. 3.13–3.17. The

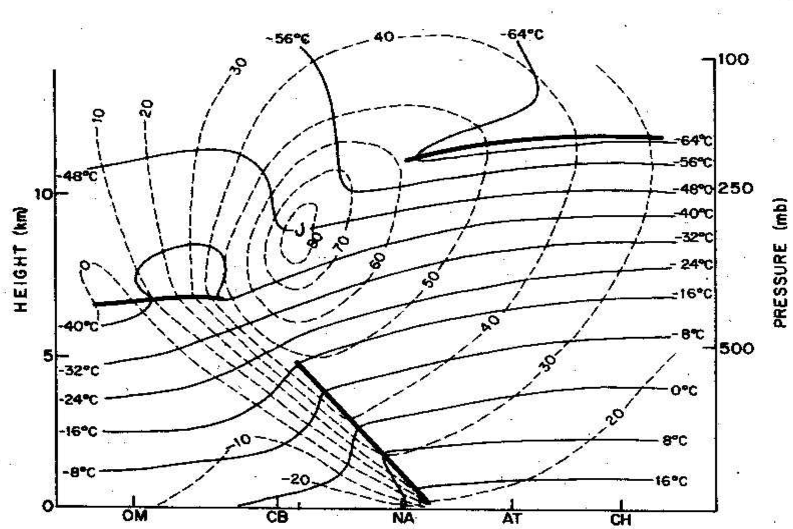

Fig. 3.19 Distribution of isotherms in degrees Celsius (——) and isotachs in meters per second (– – –) in a vertical cross section through the cold front at 00 GCT 20 November 1964. Station locations are indicated in Fig. 3.9. Isotachs refer to the geostrophic wind component normal to the section. Positive values indicate winds directed into the section. The heavy lines indicate the cold front and the tropopause. J refers to the axis of the jet stream.

isotachs (lines of constant wind speed) in the cross section refer to the wind component normal to the section. The sign convention is such that "positive" velocities refer to winds directed into the section, and "negative" velocities refer to winds directed out of the section.[†] The section is oriented approximately normal to the front and to the jet stream. The following features are evident in the figure:

- There exists a well-defined frontal zone in the lower troposphere, sloping toward the cold air with increasing height. In agreement with Fig. 3.18, the front does not extend as far west as CB.
- Within the frontal zone the isotachs are sloping and very close together, which indicates that the wind component directed into the section is increasing very rapidly with height. Such a region is said to be characterized by strong *vertical wind shear* (literally, a large vertical gradient of the horizontal wind vector).
- The middle-latitude tropospheric jet stream is located within the gap in the tropopause. As noted in the discussion of Fig. 3.16, the 250-mb surface

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[†] The isotach analysis is based not upon actual winds but upon geostrophic winds, as defined in Section 8.4.1. For the purposes of this qualitative discussion the distinction between geostrophic winds and actual winds is unimportant.

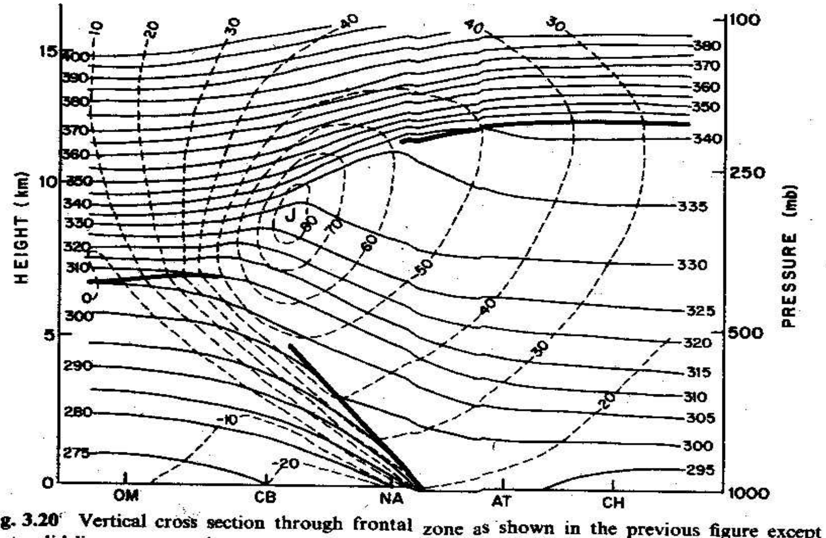

Fig. 3.20 Vertical cross section through frontal zone as shown in the previous figure except that solid lines represent isentropes (lines of constant potential temperature), labeled in degrees Kelvin.

crosses from troposphere to stratosphere at the jet stream, which crosses through the section near CB. From the section it is evident that vertical soundings taken in the vicinity of the jet stream do not show a well-defined tropopause.

- There is a reversal in the horizontal temperature gradient between troposphere and stratosphere.

In vertical cross sections it is sometimes convenient to display the distribution of potential temperature, rather than temperature. Under adiabatic conditions the isentropes (potential temperature lines) in such sections can be closely identified with air motions. In Fig. 3.20 we have reanalyzed the section shown in the previous figure in terms of isentropes and isotachs. In such analyses the stability stratification is directly related to the vertical spacing of the isentropes. Regions of close spacing (for example, the stratosphere and the frontal zone) are characterized by strong static stability.

3.5 THICKNESS AND ITS RELATIONSHIP TO VERTICAL STRUCTURE

Given the distribution of geopotential height on any two pressure surfaces, it is possible to deduce the distribution of thickness for the intervening layer. If the height fields for the two pressure surfaces are analyzed using contour lines that are integral multiples of the same contour interval (for example, 60 m) and superimposed upon the same geographical grid, the corresponding thickness chart can be derived conveniently through the procedure of "graphical subtraction." This procedure makes use of the fact that the thickness is known at each of the intersection points between the two sets of height contours; at each intersection point it is simply the difference between the labels on the two intersecting contour lines. It is clear that the value of the thickness at each of the points where the upper and lower contours intersect must be divisible by the standard contour interval. Thus the family of thickness contours that correspond to integral multiples of the standard contour interval must pass through the intersection points of the upper and lower height contours. It is also clear that one of these thickness contours cannot cross a height contour except at points where the upper and lower height contours intersect. Hence, when the three sets of contours are superimposed, all intersections must be three-way intersections.

3.5.1 The 1000–500-mb layer

Figure 3.21 shows the distribution of thickness for the 1000–500-mb layer at 00 GCT 20 November 1964. Superimposed upon the same analysis are the 500-mb height contours (heavy solid lines) and the 1000-mb height contours (lighter solid lines). The latter set of contours is derived directly from the sea level pressure analysis, using the approximation that the pressure drops 1 mb for each 8 m of vertical ascent. Thus, to within an accuracy of better than 10% the 1000-mb isobar can be relabeled as the zero height contour for the 1000-mb surface, the 1008-mb isobar as the 60-m height contour, the 992-mb isobar as the −60-m height contour, and so on. It is evident that all intersections in Fig. 3.21 are three-way intersections. By referring to the appropriate labels on the lines, it can be readily verified that at any of these intersection points, the 1000-mb height plus the 1000–500-mb thickness, is equal to the 500-mb height. Following along any given 500-mb height contour through several successive intersection points, it can be seen that, if the 1000-mb height increases by 60 m, the thickness decreases by the same amount, and vice versa, so that the sum of the two remains constant. Thus as one approaches a surface low the thickness contours cut across the 500-mb height contours toward lower 500-mb height and vice versa.

In view of the proportionality between thickness and virtual temperature, as discussed in Section 2.2.3, it is reassuring to find that the 1000–500-mb thickness distribution shown in Fig. 3.21 resembles the temperature distributions for the 850-, 700-, and 500-mb levels as displayed in Figs. 3.13–3.15. The vertical averaging inherent in the thickness field tends to smear out the frontal zones somewhat, but they are still evident as regions of strong thickness contrast on the cold side of the surface fronts. Since the 1000-mb surface is rather flat in comparison to the 500-mb surface, there is a strong correspondence

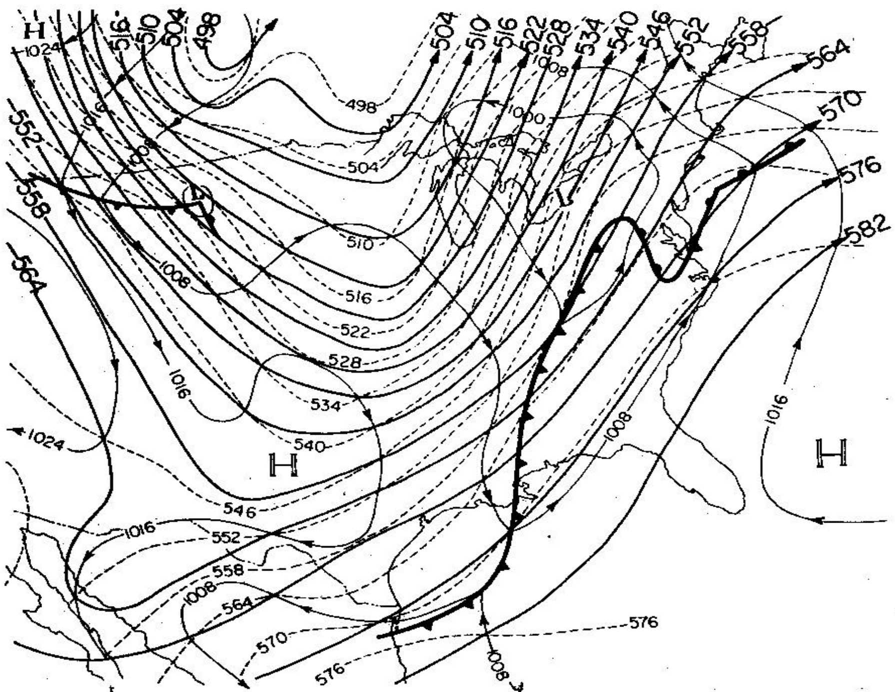

Fig. 3.21 Distribution of sea level pressure (——), 500-mb height (——), and 1000–500-mb thickness (– – –) at 00 GCT 20 November 1964. Thickness and height contours are labeled in tens of meters. Arrows on contours denote the direction of the geostrophic wind. Letters H and L refer to maxima and minima the sea level pressure field.

between the lower tropospheric temperature field and the 500-mb height field. For example, it is evident that the deep 500-mb low over central Canada is largely a reflection of the low 1000–500-mb thicknesses associated with the deep pool of cold air which covers that region.

The northerly flow behind the surface low over the eastern Great Lakes has carried some of the cold Canadian air southward into the central United States, giving rise to a pronounced "trough" in the thickness field over that region. It is because of this trough in the thickness field that the 500-mb trough is located well to the west of the surface low. In a similar manner, the cold air outbreak following behind the surface low over North Dakota shows up as a trough in the thickness field, which is responsible for the 500-mb trough upstream from the position of that surface low. The general tendency for middle-latitude disturbances to slope westward with height in the lower troposphere is thus seen to be a consequence of the distortion of the temperature field by the lower tropospheric flow patterns, which tend to bring colder air equatorward behind the surface lows and warm air poleward ahead of them.

Figure 3.22 shows a model of an idealized middle-latitude (northern hemisphere) disturbance in three stages of development, with the 1000- and 500-mb height fields and the thickness field superimposed as in the previous figure. In the initial chart the surface low is just beginning to form as a wave along a front on the warm side of the region of strong thickness contrast. In the second chart cold air is streaming southward behind the surface low and warm air is advancing northward ahead of it. These distortions in the temperature field are reflected in the growing amplitude of the "wave" in the thickness pattern. Note also how the thickness pattern is closely related to the position of the warm and cold fronts. In this stage of development the surface low is in the process of passing under the jet stream, from the warm to the cold side. In the third chart the occlusion process has begun and the surface low has begun to move across the thickness contours toward lower values as it deepens progressively farther back into the cold air. The junction of the warm and cold fronts remains on the warm side of the region of strong thickness contrast and the occluded front coincides with a warm "ridge" in the thickness field. As the disturbance continues to amplify, the positions of the surface low and the 500-mb trough (or closed low) gradually begin to come into vertical alignment. In fully developed systems the vertical tilt completely disappears and all three sets of contours become mutually parallel.

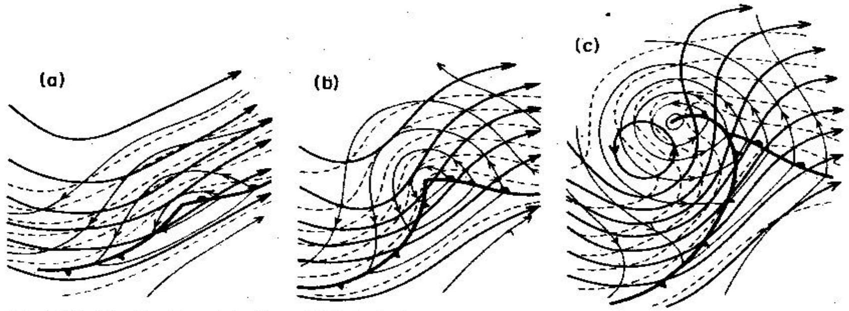

Fig. 3.22 Idealized model of a middle-latitude storm in three stages of development: (a) initial stage, (b) developing stage, and (c) occluded stage. (——) isobars of sea level pressure, (——) contours of 500-mb height, (– – –) contours of 1000–500-mb thickness. (From E. Palmén and C. W. Newton, "Atmospheric Circulation Systems," Academic Press, New York, 1969, p. 326.)

3.5.2 The 250–100-mb layer

The distribution of temperature in the 250–100-mb layer at 00 GCT 20 November 1964 is displayed in Fig. 3.23. The thickness pattern qualitatively resembles the 100-mb temperature field shown in Fig. 3.17, and it is remarkably similar to the 1000–500-mb thickness pattern except for the fact that the

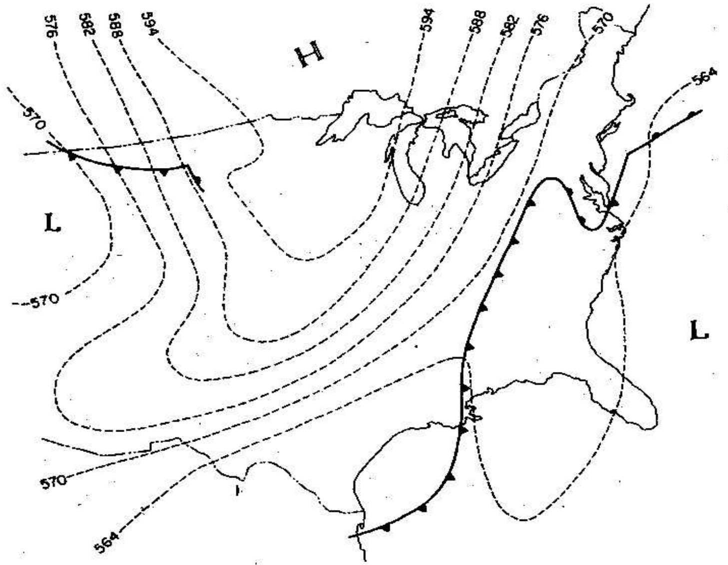

Fig. 3.23 Distribution of 250–100-mb thickness at 00 GCT 20 November 1964. Thickness contours are labeled in tens of meters. Frontal positions are transcribed from the surface chart.

gradients are reversed. The tendency for compensation between tropospheric and lower stratospheric temperature fields was remarked upon in the previous section. We may interpret this tendency for compensation as follows. Immediately above the tropopause the temperature field reverses so that troughs (or closed lows) become warm relative to their surroundings and ridges become cold. The resulting out of phase relationship between thickness and geopotential height is consistent with a reduction in the amplitude of synoptic-scale features as we proceed upward to pressure levels higher in the stratosphere. Troughs tend to be filled in by the positive thickness anomalies and ridges tend to be flattened out by the negative thickness anomalies. As the pressure surfaces become progressively flatter, the wind speeds decrease with height. By the time we reach the 100-mb level, there is little remaining evidence of synoptic-scale features in the geopotential height field. The planetary-scale features present at that level and above have a vertical structure much different from the disturbance that we have considered in this section.

Thus the tendency for compensation between tropospheric and lower stratospheric temperature anomalies is seen to be intimately related to the characteristic vertical structure of a broad class of middle-latitude synoptic-scale disturbances, which exhibit their maximum amplitudes at the tropopause level,

and damp out rapidly with height in the lower stratosphere. Unlike the tropospheric temperature gradients, which own their existence to strong latitudinal and geographical contrasts in diabatic heating near the ground, the stratospheric temperature gradients are induced adiabatically by vertical motions in the vicinity of the tropopause. Sinking motion induces warming and suppresses the height of the tropopause above domes of cold tropospheric air, while rising motion induces cooling and lifts the tropopause above the warm tropospheric air masses. We will consider the role of vertical motions in middle-latitude disturbances in more detail in Section 9.5.

PROBLEMS

3.1 Explain or interpret the following:

(a) The temperatures observed at cold fronts are often as high as those observed anywhere within the warm sector of a storm.

(b) Temperature falls that follow the passage of cold fronts are usually more rapid than the temperature rises that precede the passage of warm fronts.

(c) Temperature falls that follow the passage of cold fronts are usually most pronounced in situations where the frontal passage takes place during the late afternoon or evening.

(d) Pressure tendencies are not much help in locating stationary fronts.

(e) Fronts over the sea show up much more clearly on the 850-mb map than on the surface map.

(f) In mountainous areas the winds bear little relation to the isobars on the surface map.

(g) The combination of high sea level pressure and low 500-mb height is usually accompanied by below normal temperature.

(h) Large pressure and temperature changes occur prior to the passage of a warm front, but after the passage of a cold front.

(i) On the average, occluded fronts tend to be weaker than warm or cold fronts.

(j) Soundings taken in the vicinity of the jet stream do not exhibit a well-defined tropopause.

(k) The presence of a stable layer in a sounding may or may not be an indication that a front is present. What other information in the sounding can be used to determine whether a front is present?

(l) The northern boundary of the Gulf Stream is a favored region for the development of winter storms.

(m) In the northern hemisphere, the passage of either a warm or a cold front produces a windshift in the clockwise sense. In what sense are the windshifts in the southern hemisphere?

(n) Freezing rain is frequently observed in association with warm fronts, but rarely with cold fronts.

(o) Middle-latitude disturbances usually tilt westward with height.

3.2 (a) For the idealized frontal configuration shown in the accompanying figure, show that the slope of the front is given by $dz/dx = \beta/(\Gamma_w - \Gamma_F)$, where Γ_w is the lapse rate on the warm side of the front, Γ_F is the lapse rate within the frontal zone, and β is the absolute magnitude of the horizontal temperature gradient within the frontal zone. [Hint: Express the temperature difference between points A and D in terms of the gradients encountered along the path ABD and along the path ACD, and equate them.]

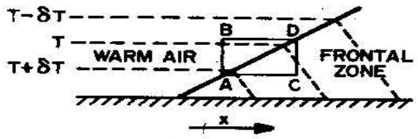

Fig. P3.2.

(b) Given a lapse rate of 7 deg km^{-1} in the warm air and zero in the frontal zone, and a horizontal temperature gradient of 10 deg per 100 km within the frontal zone, calculate the slope of the front.

Answer 14.3 m km^{-1}

3.3 (a) Show that for the idealized frontal zone shown in the figure the mean temperature of the layer BB′ is lower than that of the layer AA′ by the amount $\frac{1}{2}\beta x_B$, where β is the absolute magnitude of the horizontal temperature gradient within the frontal zone.

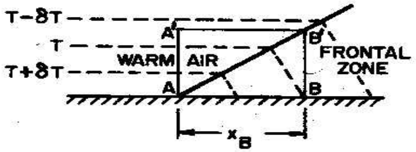

Fig. P3.3.

(b) If the pressures and altitudes at points A′ and B′ are equal, prove that the ratio of the pressure at point B to that at point A is given by

$$\frac{p_B}{p_A} = \left(\frac{p_A}{p_{A'}}\right)^{(T_A - T_B)/T_B}$$

where p_A, p_B, and $p_{A'}$ are the pressures at points A, B, and A′, respectively, and T_A and T_B are the mean virtual temperatures of the layers AA′ and BB′, respectively.

3.4 On the basis of the expression derived in the previous problem, estimate the slope of the stationary front shown in Figs. 3.3a and 3.4a. (200 km to the north of this front the surface temperature is lower by ≃ 10 deg and the sea level pressure is higher by ≃ 4 mb than at the surface position of the front.) Assume that $T_B \simeq 270$°K.

Answer ≃ 100 mb per 100 km