MODULE 2.5E

TEPHIGRAMS AND HODOGRAPHS

Hodograph Analysis

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Introduction

Among the observed weather elements which have great usefulness in forecasting are wind observations. However, quite apart from their immediate application to wind forecasting problems, they can, through systematic interpretation, yield valuable indirect information about the thermal structure of the atmosphere even in the absence of direct temperature measurements. Hodographs can be used to monitor:

- The speed and direction of warm and cold fronts
- Differential temperature advection , changing stability
- Expected vertical motion near fronts

The physical principle underlying the interpretation of wind observations is the geostrophic thermal wind equation, which may be written in the following two forms where the symbols have their usual meaning:

$$\frac{\partial \vec{V}_g}{\partial z} = \frac{g}{fT_p}\hat{k} \times (\nabla T)_p \qquad (1)$$

$$\vec{V}_T = \frac{g}{f}\hat{k} \times \nabla H \qquad (2)$$

Equation (1) states that the magnitude of the geostrophic vertical wind shear vector is proportional to the isobaric temperature gradient and is directed parallel to the isotherms on the isobaric surface in the sense that lower temperatures are found to its left, relative to an observer facing downwind along it. Equation (2) states that the magnitude of the total thermal wind vector, V_T for an isobaric layer is proportional to the horizontal thickness gradient of that isobaric layer and is directed parallel to the thickness lines in the sense that lower thicknesses are found to its left. Since thickness lines are equivalent to mean isotherms of the layer, it may also be stated that V_T is directed parallel to the mean isotherms of the layer in the sense that lower temperatures are found to its left. The plane geometric relations expressed by equations (1) and (2) are shown in Figure 1.

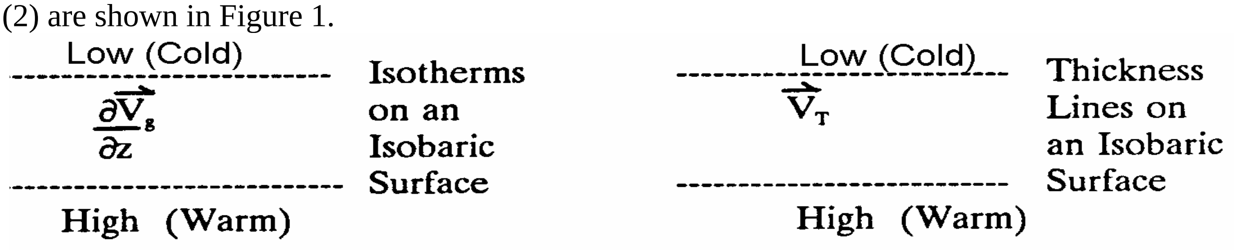

Figure 1. The thermal wind.

In summary: the stronger the horizontal temperature gradient, the stronger the vertical wind shear; and the stronger the horizontal thickness gradient the stronger the thermal wind.

Plotting the Wind Shear on the Hodograph Form

Representation of Winds

To analyze upper wind soundings, the winds are plotted on polar coordinate paper. The forms are called wind shear hodograph forms or, simply, hodograph forms. The coordinates are wind speed (V) and angle (r), respectively. The radial lines are labelled in degrees while the concentric circles are labelled in speed as shown in Figure 2.

A wind at any level is represented by a point on this diagram corresponding to the end point of the wind vector directed radially outward from the centre together with the height for which the wind was determined. The magnitude of the wind speed is given by the length of the vector measured on the speed scale. To give this diagram some physical reality, one can imagine a greatly enlarged hodograph form mapped on the earth's surface centered on the observing station, with the 180° radius directed northward. The wind at any level may then be visualized to blow away from the observer along the wind vector.

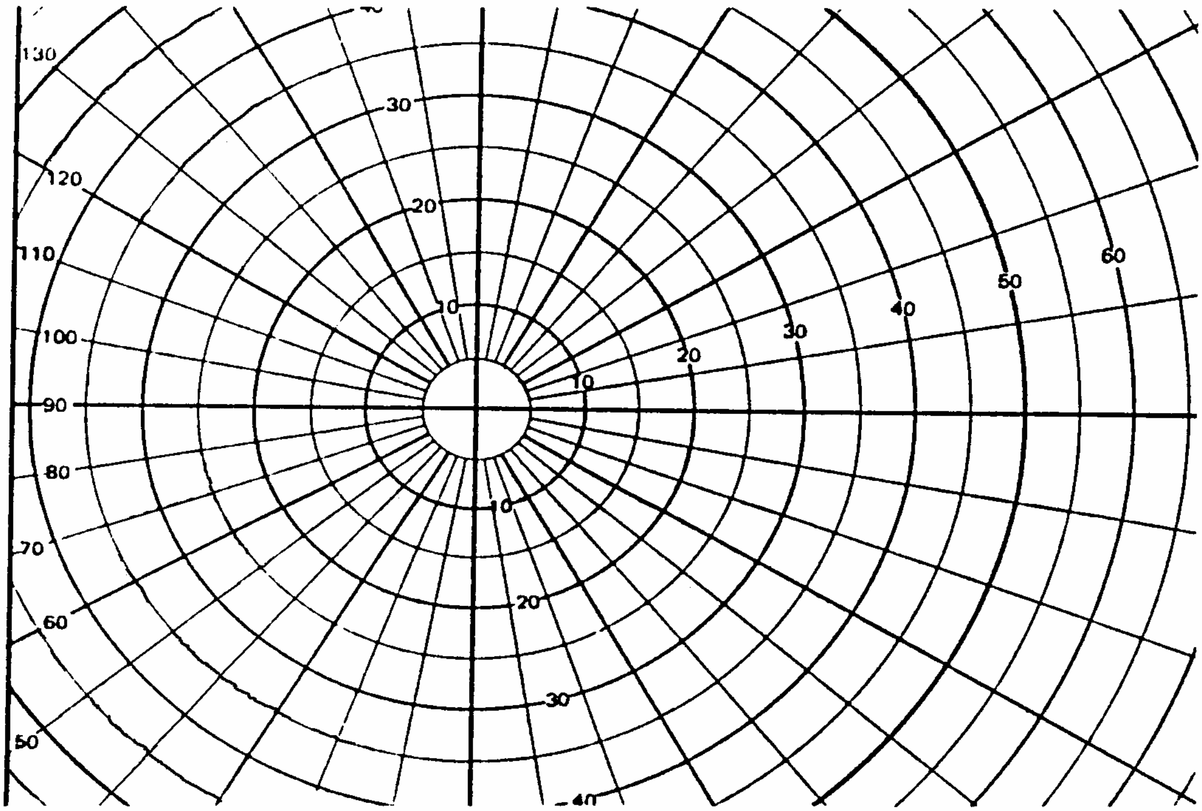

Figure 2. The Hodograph.

When winds change from south to east, they are said to "BACK from 180 to 90 degrees". (Mnemonic: back means to count backwards). Winds increasing their direction from 180 to 270 are said to "VEER".

The Total Thermal Wind and Vertical Wind Shear

Consider winds for two different levels plotted on a hodograph form. The wind for the lower level is represented by the vector from the origin to the plotted points. The same is true for the wind for the higher level.

The Total Thermal Wind Vector, V_T

The total thermal wind vector $\mathbf{V}_T$ is the vector difference between the wind vectors for the upper and lower levels. It is designated by the usual wind notation, that is, the magnitude of $\mathbf{V}_T$ is given by its length measured on the speed scale, while its direction is denoted by the direction of the nearest parallel radius on the hodograph form having the same sense as $\mathbf{V}_T$. Thus, if the 5,000-ft and 10,000-ft winds were respectively 270/30 and 360/40 as plotted in Figure 3, the thermal wind vector for the layer would be denoted 040/50.

The Vertical Wind Shear, $V_T/\Delta Z$

The vertical wind shear is the average wind shear through the layer as defined by the ratio $\mathbf{V}_T/\Delta Z$. It is a scalar quantity where $\mathbf{V}_T$ is the magnitude of the thermal wind vector and ΔZ is the height difference typically expressed in thousands of feet.

In Figure 3, $\mathbf{V}_T$ has a value of 50 knots over a height difference of 5000 feet. The vertical wind shear is thus (50/5) kts/1000 ft. The value of the shear is correlated with the turbulence experienced by an aircraft.

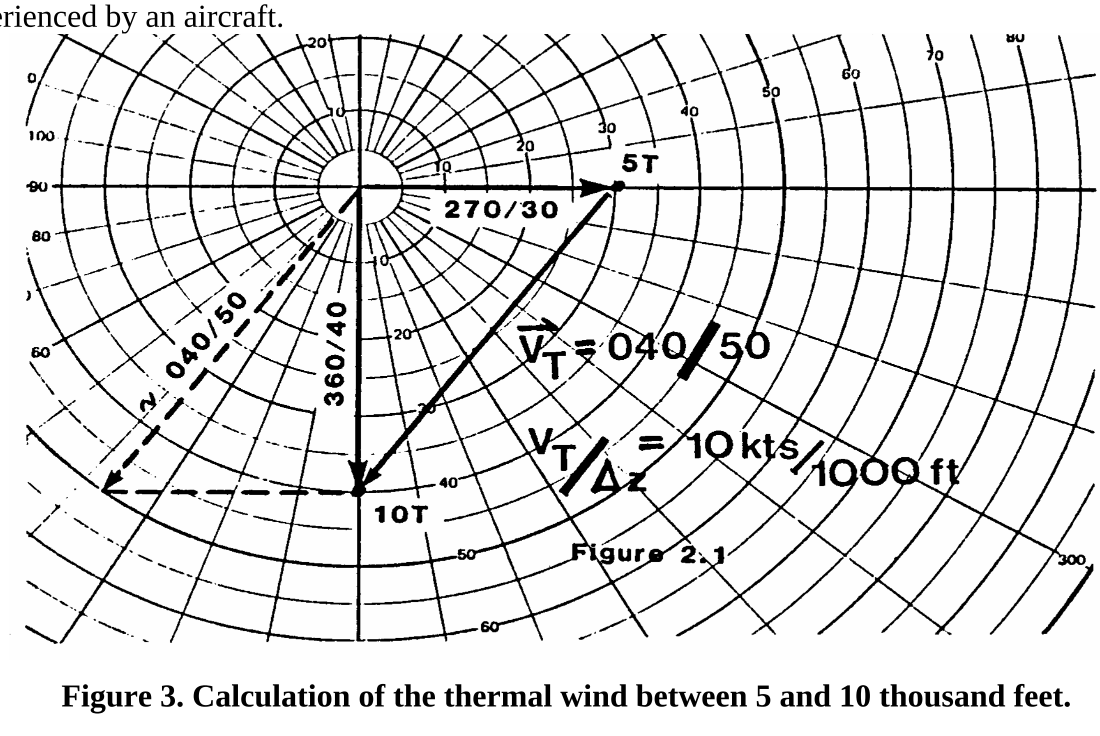

Figure 3. Calculation of the thermal wind between 5 and 10 thousand feet.

Analysis of hodographs

Relation Between the Thermal Wind Vector and Horizontal Temperature Gradient

The geometric relation between the geostrophic thermal wind vector $\mathbf{V}_T$ for a layer of the atmosphere, the geostrophic winds at the base and top of that layer, and the thickness lines or mean isotherms of that layer is shown below.

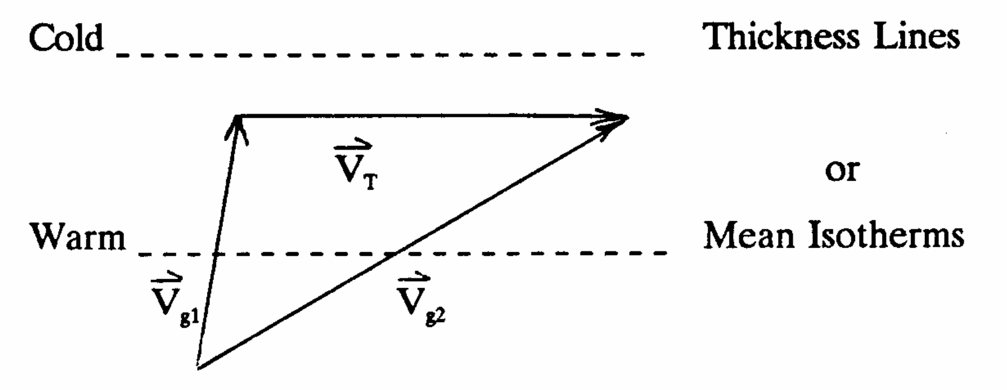

Figure 4. The Thermal Wind Vector.

In the above example, the mean isotherms in the layer of the atmosphere associated with the thermal wind vector $\mathbf{V}_T$ will be oriented west to east with colder air to the north of the station. Just like the geostrophic wind blowing with lower pressure to the left, here is a simple mnemonic for the thermal wind:

If you stand with the thermal wind at your back, the cold air is on your left.

The simple relation between the thermal wind vector and the mean temperatures is acceptable only when the actual winds are nearly geostrophic. Operationally, the actual thermal wind is identified with the geostrophic thermal wind. In practice, this means that it is valid only in the free atmosphere above the friction layer, (say, higher than 3,000 ft above the earth's surface) and when there are no appreciable accelerations or vertical motions present. Further, the relation is not accurate in highly curved flows. Despite these drawbacks, the thermal wind is a powerful forecasting tool.

Height of the Gradient Wind Level

A study of the shear hodograph will give an indication of the depth of the friction layer, or PBL (Planetary Boundary Layer). The wind in the PBL is a combination of the gradient wind (geostrophic + curvature), friction effects and the horizontal temperature gradient. Frictional interaction of the atmosphere with the underlying surface produces a characteristic veering and increase of wind with height known as the Ekman spiral (Figure 5).

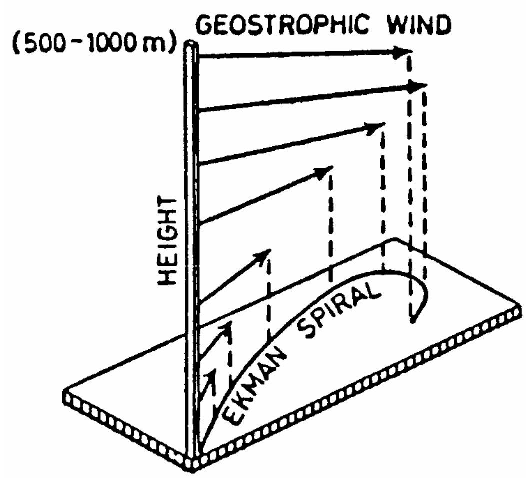

Figure 5. The Ekman Spiral.

The vertical wind shear associated with horizontal temperature gradients is additive to the Ekman spiral shear. It is difficult to diagnose the wind shear in the PBL resulting solely from temperature gradients.

The top of the PBL is identified at the point where there are no more frictional effects. Above the PBL the wind is a result of gradient balance or nearly geostrophic. Typically at the top of the PBL, the wind speed increases sharply and the characteristic veering with height ends. Figure 6 illustrates the location of the top of the PBL in a couple of cases. Some situations will be more questionable.

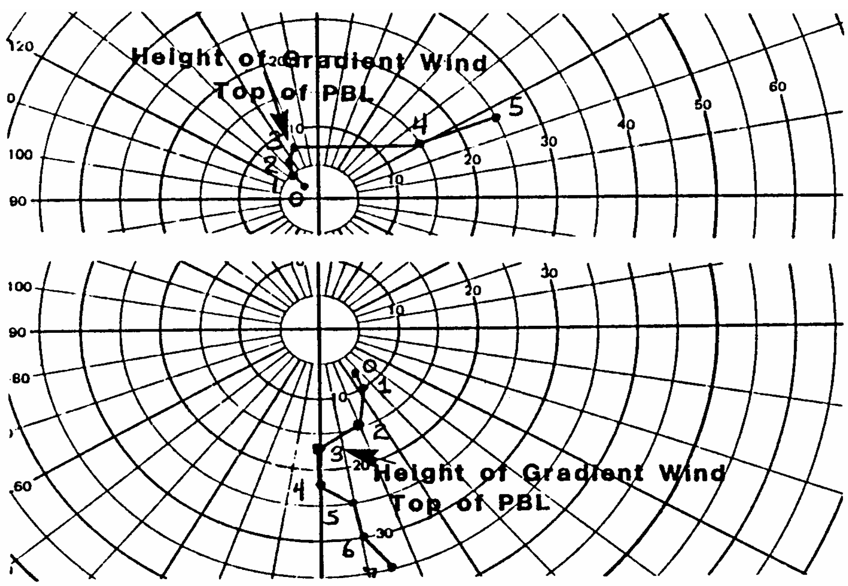

Figure 6. The Top of the PBL.

Horizontal Temperature Advection

Consider the schematic hodograph in Figure 7 in which the wind at one level is represented by OA and the wind for a higher level by OB:

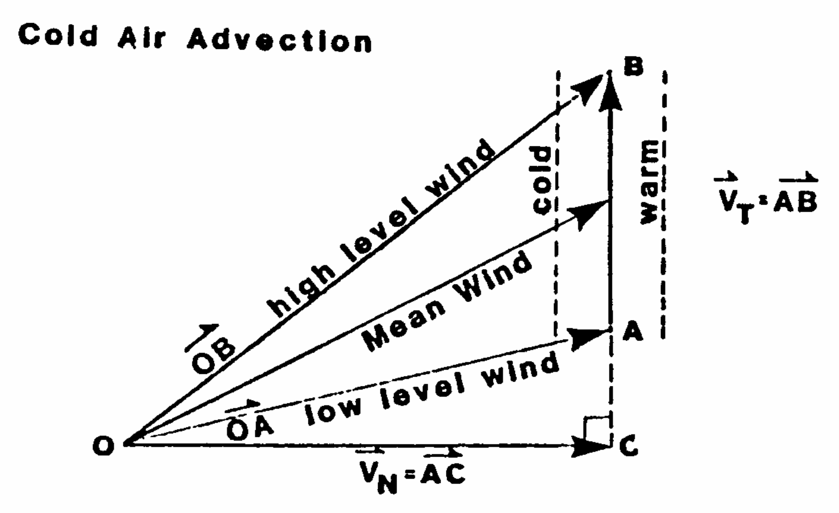

Figure 7. Horizontal Temperature Advection.

The thermal wind vector for the layer is AB. The mean isotherms in the layer will be parallel to AB with colder air toward the west. If there were a uniform variation of wind with height, the mean wind in the layer would be given by the wind vector directed to the mid-point of AB. The mean wind can be resolved into components parallel and perpendicular to AB. If the mean isotherms were regarded as material lines embedded in the mean wind field and were moved by the wind, their rate of motion would be given by the component of the mean wind normal to AB, i.e., V_N, which is found by drawing a perpendicular line from the origin of the hodograph to V_T or its projection. In the example shown in Figure 7, since the temperature in the layer will decrease with time, we say that there is cold air advection in that layer.

Consideration of warm and cold air advection will show:

- Warm air advection implies veering winds with height (see Fig. 6)
- Cold air advection implies backing winds with height (see Fig. 7)
- No temperature advection implies no change in wind direction

Mnemonic: **B.C. VolksWagon**:

Backing - Cold air advection and Veering , Warm air advection.

Magnitude of the Horizontal Temperature Advection

Meteorologists are interested in the strength of the temperature advections at various levels because it tells them how quickly the atmosphere is stabilizing or de-stabilizing. This in turn affects our forecasts of convection, its intensity, and duration. We will be able to answer the question of whether things will get stronger, weaker, or stay the same.

We can readily compute the temperature advection at a level from a hodograph. If we consider a purely advective case, with no change in baroclinicity with time, then equation 3 applies:

$$\frac{dT}{dt} = \frac{\partial T}{\partial t} + \vec{V}_N \bullet \nabla T = 0 \qquad (3)$$

$\mathbf{V}_N$ is the component of the mean wind normal to the temperature gradient. $\mathbf{V}_N$ is determined from the hodograph as in Figure 7. The temperature gradient is related to strength of the thermal wind (directly measured on the hodograph) through the thermal wind equation. With a little mathematics, we can get the temperature gradient itself and the local rate of temperature change in °C/hr (see the section on Horizontal Temperature Advection and Gradients near the end of this document).

Although the value of $\partial T/\partial t$ is extremely handy, there are some problems with it. The first is that the hodograph is an instantaneous measure of the temperature advection, and so is not always representative of a 12 hour period. Secondly, small changes in direction and speed of the wind can lead to significant errors (up to 1°C/hr). Most importantly, temperature changes in the atmosphere may also arise from diabatic processes, or vertical (adiabatic) motion. The latter two may overwhelm the advective temperature change. Proper use of the temperature advection measure requires a thoughtful analysis.

We can still make useful convective forecasts with a qualitative advection guide. The best we can hope to expect operationally, without computers or calculators, is the knowledge of how the advections are changing the stability: positively or negatively. Often, all that is required is cooling or warming by one or two degrees to set off convection.

Differential Temperature Advection and Changes in Vertical Stability

By comparing the advective rates of temperature change for two different levels, one can arrive at some qualitative inference regarding the changes in lapse rate and hence, vertical stability with time. For example, if there were a greater rate of warming at one level relative to a lower level, the lapse rate will indicate increasing stability. On the other hand, if greater advective cooling is occurring at higher levels, the lapse rate will indicate decreasing stability. The examples in Figure 8 will serve to illustrate the principles involved. For purely heuristic reasons, they are arranged to indicate the same wind component normal to the isotherms. Normally this will not be true in real situations. Remember that the greater the shear between two levels, the stronger the temperature gradient.

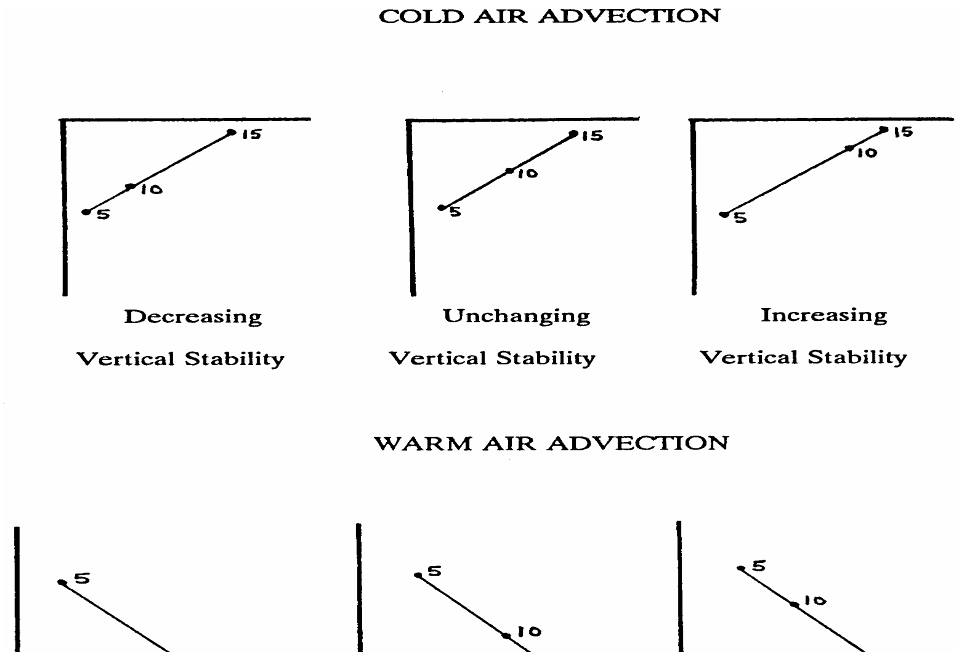

Figure 8. Stability tendency on a Hodograph.

A frequent situation in which hodograph analysis can assist in forecasting thunderstorm development is one in which cold air advection at one level surmounts warm air advection at a lower level. For example, consider the schematic hodograph in Figure 9 which appears often in the warm air ahead of a surface cold front.

Slight cold air advection (backing) between 8-12,000 ft is occurring over warm air advection (veering) between 4-8,000 ft. If the conditions of moisture and stability are suitable for thunderstorm development, the indications of decreasing vertical stability given by such a hodograph may serve to confirm the basis for forecasting thunderstorms. In fact, the cold air advection might be considered a trigger in setting them off.

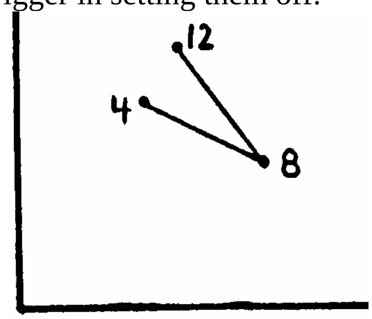

Figure 9. Cold Advection over Warm Advection.

NOTE: It is important to have a copy of the corresponding tephigram next to the hodograph. It does not really matter if there is lots of temperature advection at a particular level. The analyst must consider how that advection will affect the current state, relative to the air above and below that level. Going back to the analogy of a ball in a depression next to a cliff, if the depression is a deep pit, then even a swift quick will not do much to get the ball falling over the cliff. It is all relative.

Location of the Tropopause

The height at which the wind shear vector reverses direction can often be used to identify the tropopause, since the stratospheric temperature gradient is normally opposed to that in the troposphere. However, since winds do not always decrease with height in the stratosphere, especially in the region north of the warm band in the lower stratosphere in winter, the identification of a tropopause at the level where the thermal wind vector reverses direction is not always valid. There is a high correlation between the height of the tropopause and the temperatures at the tropopause, thus, the direction of the wind shear vector in the vicinity of the tropopause will often yield the approximate orientation of the tropopause surface. Clearly, the tephigram is more appropriate than the hodograph as a source of data for tropopause determination although the hodograph can also reveal different information as indicated.

Non-Frontal Inversions

In wintertime, the presence of a shallow radiation inversion in continental arctic air may be associated with a surface calm while the full gradient wind may be found a few hundred feet above the surface. The presence of a large vertical wind shear in this situation may not normally be interpreted in terms of a corresponding horizontal temperature gradient. See Figure 10. A quick look at the corresponding tephigram will show whether or not there is an inversion.

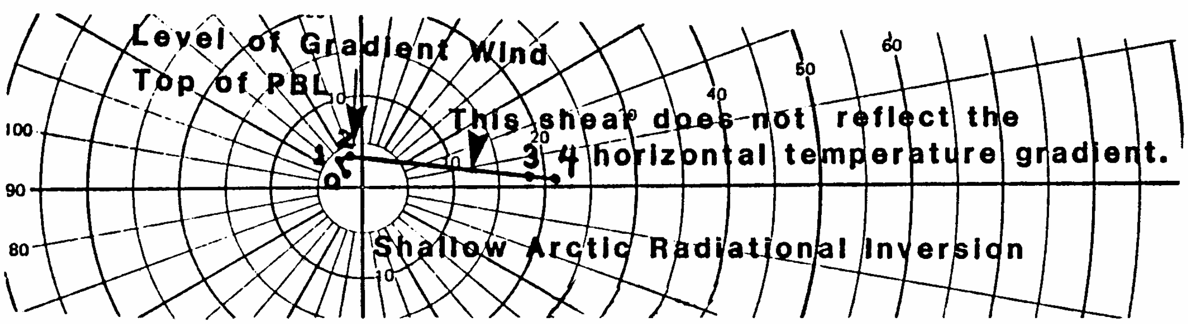

Figure 10. Shallow cold air radiation inversion.

Interpretations of Hodographs

Principles

The geostrophic thermal wind equation is the basis for the identification of fronts from wind shear hodographs. Since maximum horizontal temperature gradients occur in frontal zones, they will normally be associated with maximum vertical wind shears.

An air mass is normally characterized by relatively small horizontal temperature gradients. Consequently, the vertical wind shears in air masses will be relatively small, usually less than 2 - 4 knots per thousand feet.

A wind shear hodograph permits one to interpret the following characteristics of fronts:
1. Height of a frontal surface, base of mixing zone.

2. Orientation of frontal zone.
3. Direction of frontal motion.
4. Speed of a front.
5. Instantaneous changes in vertical stability.
6. Vertical motion in the warm air mass.

Identification of Fronts

If a hodograph exhibits a relative maximum of vertical wind shear in a certain layer, we may interpret that layer to correspond to the mixing zone of a front. Remember the schematic of warm uniform air mass overlying the colder air beneath. It is in the cold air that the gradients of the temperature and humidity lie. In particular, the top of the layer will correspond to the top of the mixing zone. In the warm air above, the temperature gradient is weak, and therefore the wind shear is weak. The hodograph analysis should be done simultaneously with the tephigram analysis to ensure a consistent interpretation in space and time.

Orientation of Frontal Surfaces

It is observed that isotherms in the mixing zone on upper-level charts are approximately parallel to the fronts at that level. The direction of the thermal wind vector in the mixing zone will therefore approximate the orientation of the frontal surface at that level. This information facilitates drawing fronts in regions of sparse data. It also simplifies the checking of the heights of fronts inferred from the tephigram analysis.

Direction of Frontal Motion

The direction of motion of a front is given by V_N the line from the origin of the hodograph perpendicular to the thermal wind vector associated with the front. Whether it is a cold front, warm front, or stationary front at that level will depend on the advection pattern. Thus, in a frontal layer, if the wind backs with height, it will be a cold front, if it veers, it will be a warm front; finally, if it neither backs or veers (the shear is oriented along a radial on the hodograph), it will be stationary. Be careful not to choose the value 180 degrees out. Check against a synoptic map.

Speed of a Front

The speed of a front at the level identified on a hodograph will be the magnitude of V_N measured on the speed scale. In the absence of vertical motions, this will correspond very closely to its instantaneous speed. This is very valuable for short range prognosis. It serves as an excellent check for frontal motion.

Instantaneous Change in Vertical Stability

The instantaneous changes in vertical stability in the vicinity of a station may be assessed using the principles outlined in a previous section. This analysis is particularly useful if it is made in conjunction with a stability analysis of the corresponding tephigram.

Vertical Motion in the Warm Air Mass

Shear hodographs can frequently indicate indirectly the field of vertical motion in the warm air mass above a frontal surface. This knowledge may often serve as the basis for short-range cloud forecasting. The interpretation of the field of vertical motion from hodographs is based on one assumption and one principle. The assumption is that the frontal surface does not change slope for a short period of time. The principle is the equation of continuity, which implies that horizontal divergence must be accompanied by vertical motion.

Consider a vertical cross section in a plane perpendicular to a cold front, where the point P indicates the point of observation. The speed of the front at 8,000 feet is represented by the arrow. If the frontal surface is assumed to move at the same speed at a higher level, and if the component of the warm air wind perpendicular to the front decreases with height, there is horizontal convergence ahead of the front, and consequently, ascent in the warm air. If the warm air mass is potentially unstable, the lift may be sufficient to realize the instability. It is therefore said that cold fronts showing this characteristic wind field are active. One can, therefore, formulate the following rule: if the wind shear vector above a cold frontal surface backs relative to the frontal shear, there is ascent in the warm air.

Figure 11. Active (or Anabatic) Cold Front.

Figure 12 shows a schematic cross-sectional view of an active cold front. The term “active” or “inactive” refers to the presence or absence of weather on the front. In some meteorological literature, the active front is referred to as an anabatic front. The ascent in the warm air is sometimes termed “rearward sloping ascent”.

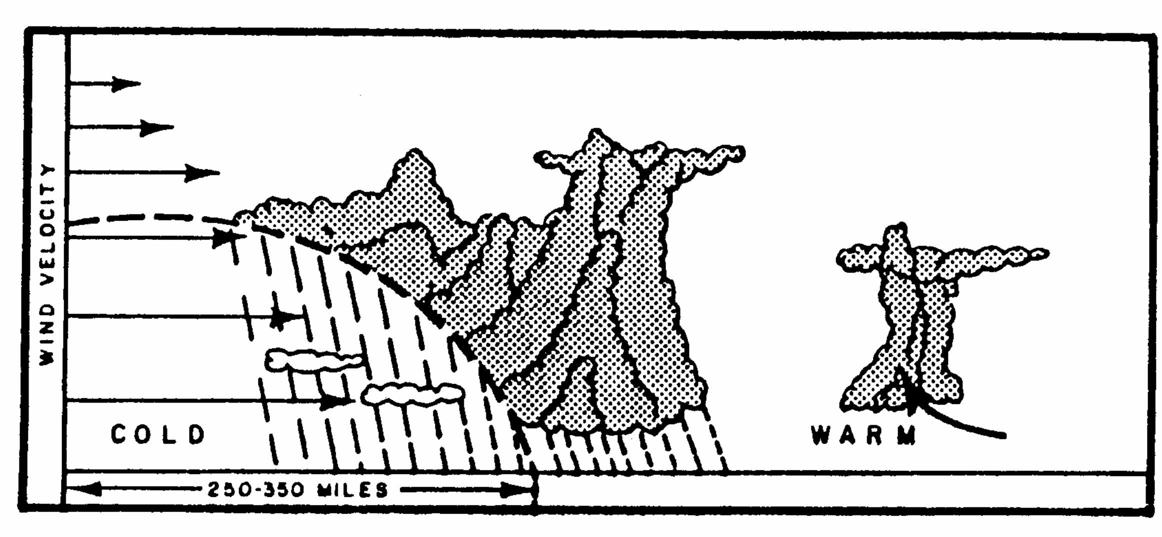

Figure 12. Cross Section of an Active or Anabatic Cold Front.

In the example in Figure 13, there is horizontal divergence ahead of the front since the component of the warm air wind perpendicular to the front is greater than the frontal speed. This must be accompanied by subsidence. Cold fronts showing this characteristic wind field are termed inactive or katabatic. One can formulate the following rule: If the wind shear vector above a cold frontal surface veers relative to the frontal shear, there is descent in the warm air.

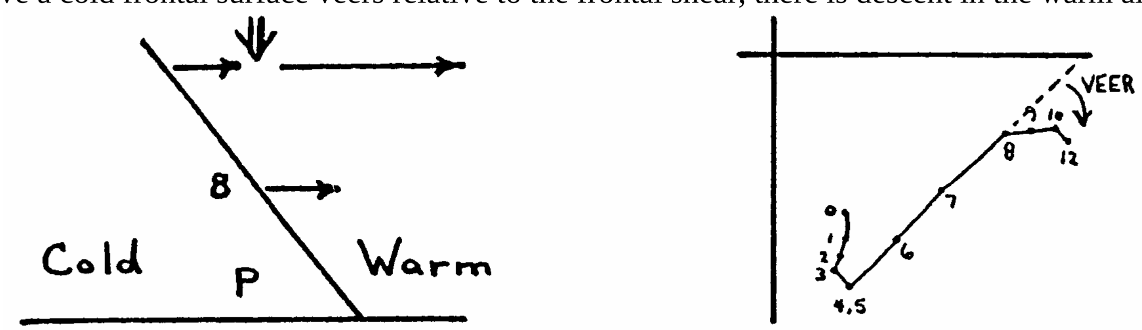

Figure 13. Inactive or Katabatic Cold Front.

Figure 14 depicts a cross-section of an inactive or katabatic surface cold front. It is important to note here that we are emphasizing that the cold front is inactive at the surface to avoid confusion with an upper cold front (or upper humidity front). As shown in the diagram, a dry cold front is accompanied by clear skies (hence inactive), but out ahead in the warm air there can be an intense line of convection supported by the moisture in the warm sector. You will see in the modules dealing with fronts that out ahead of the katabatic front there is often an upper humidity front which supports the precipitation. The distance between the inactive cold front and the upper humidity front varies from zero to a couple of hundred kilometers depending on the system's evolutionary state.

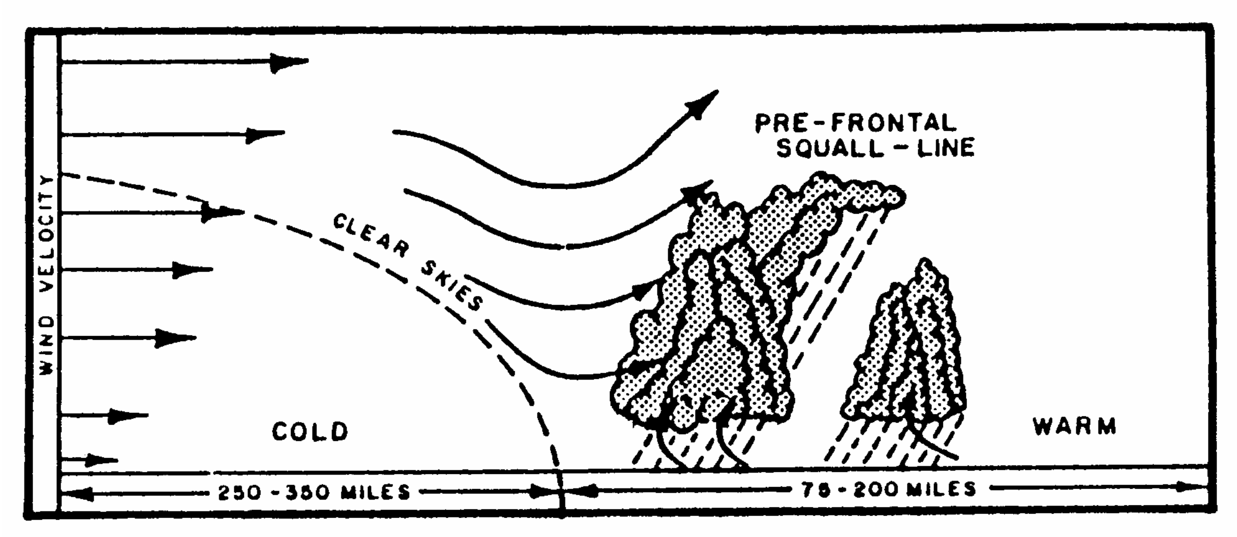

Figure 14. Cross Section of an Inactive or Katabatic Surface Cold Front.

In the example shown in Figure 15, there is horizontal convergence in the warm air that must be accompanied by ascent. Warm fronts showing this characteristic wind field are called active, because of potential cloud development. Thus, if the wind shear vector above a warm frontal surface backs relative to the frontal shear, there is ascent in the warm air.

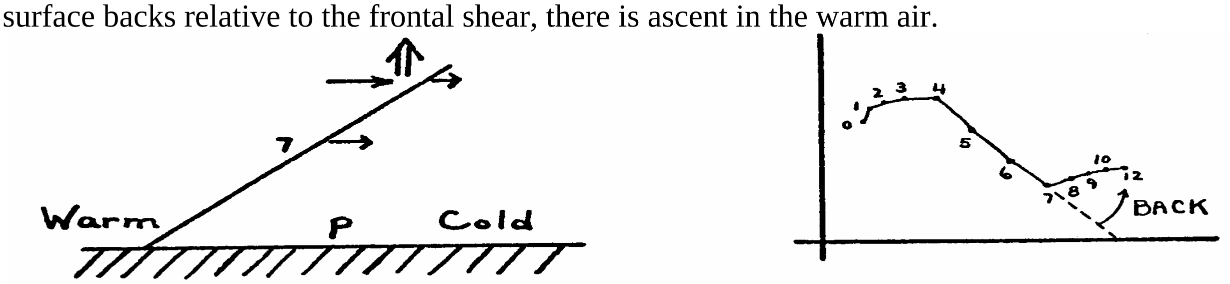

Figure 15. Active or Anabatic Warm Front.

In the example shown in Figure 16, there is horizontal divergence in the warm air since the component of the warm air wind perpendicular to the front is less than the frontal speed. This situation must be associated with subsidence. Such warm fronts are said to be inactive. Thus, if the wind shear vector above a warm frontal surface veers relative to the frontal shear, there is subsidence in the warm air.

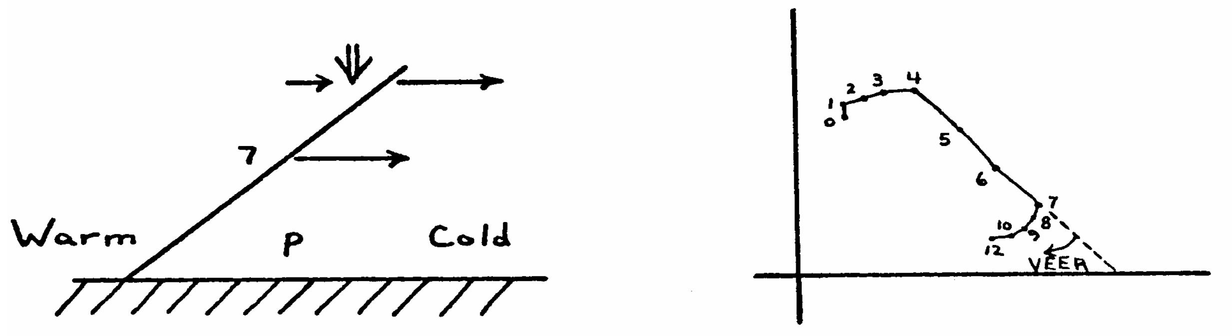

Figure 16. Inactive or Katabatic Warm Front.

Note that in the case of both warm and cold fronts, a backing wind in the warm air relative to the frontal shear denotes an active front. The converse is also true.

Hodograph Analysis Methodology

It may be useful to develop a uniform approach when learning hodograph analysis and interpretation. The following suggests one approach which starts with the easier elements and progresses to those which are more difficult.

1. Identify the Gradient Wind Level (Top of the PBL)
2. Identify the wind at the top of the PBL. (Useful for estimating the surface wind).
3. Identify layers with relatively strong vertical wind shear. For each layer, determine the:
- top (frontal surface) and base of the mixing zone
- vertical wind shear (for turbulence forecasting)
- orientation of the frontal zone
- speed and direction of motion of the front
- vertical motion in the warm air mass
4. Identify the changes in vertical stability that would result from the differences in the vertical temperature advections.
5. Collaborate the results of the hodograph analysis with other data such as tephigrams, upper air analyses, and satellite imagery.

Example of Hodograph Analysis and Interpretation

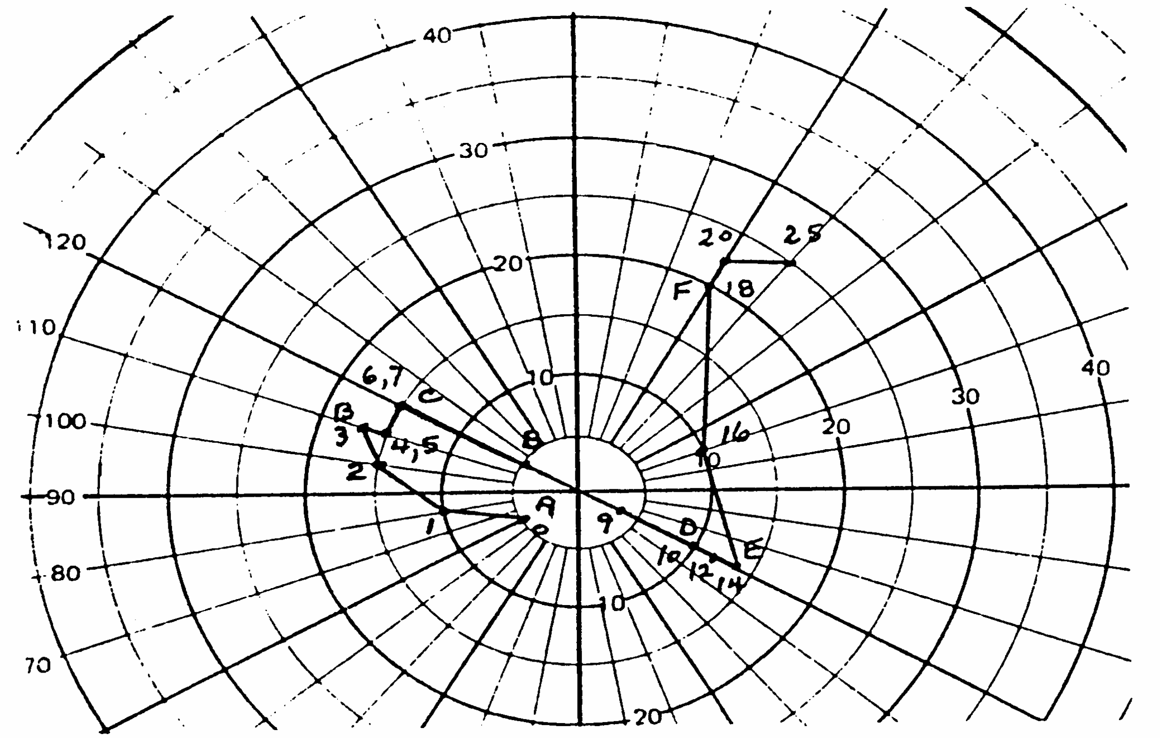

Figure 17.

The analysis and interpretation of the hodograph in Figure 17 would proceed along the following lines:

AB is the shear in the friction layer, since it shows the normal veering and increase with height.

The gradient wind level is 3,000 ft.

BC yields a very small vertical wind shear, hence it likely occurs within an air mass.

CD gives a vertical wind shear of 8 kt per 1000 ft, which is a relative maximum, hence it may identify a frontal zone. The frontal surface is at 10,000 ft while the base of the mixing zone is at 7,000 ft.

The frontal surface is oriented WNW-ESE with colder air toward the northeast of the station.

The front is stationary. (V_N is zero).

Since the components of the winds in the warm air above the frontal surface represented by DE are the same as the frontal speed, there is little vertical motion above the frontal surface.

EF is the shear in a frontal zone, since there is a relative maximum of 6 kt per 1,000 ft in the 14,000 to 18,000 ft layer. The frontal surface is at 18,000 ft while the base of the mixing zone is at 14,000 ft.

The frontal surface is oriented south-north with colder air to the west of the station.

It is a cold front moving eastward (from 270 degrees) at around 10 kt.

Since the components of the warm air winds perpendicular to the front increase with height, it indicates slight subsidence in the warm air mass.

With cold air advection between 14,000 and 18,000 ft surmounting little advection at lower levels, the vertical column of the atmosphere below 18,000 ft must show decreasing vertical stability with time.

Horizontal Temperature Advection and Gradients

The magnitude of the advection rate of temperature and the spacing of the isotherms can be computed from a hodograph. Actual temperature changes in the atmosphere may arise from diabatic processes, vertical motion or horizontal temperature advection. Only in the absence of the first two processes will the actual temperature change correspond to the advective temperature change as identified above.

The magnitude of the advection rate of temperature change can be computed from a hodograph. Consider a set of isotherms in a horizontal plane which moves with the wind. For a coordinate system fixed to a particular isotherm the temperature is constant, and hence, its total derivative with respect to time is zero (assuming no external forcing H). Now, temperature (T) is a function of time and distance normal to the isotherms, hence, the total derivative may be expanded as follows: (See Figure 18)

$$\frac{dT}{dt} = \frac{\partial T}{\partial t} + \frac{\partial T}{\partial n} \bullet \frac{d\vec{n}}{dt} + \dot{H} \qquad (4)$$

Since $\vec{V}_N = \frac{d\vec{n}}{dt}$ and $\frac{dT}{dt} = 0$, by using the finite difference form we demonstrate that the total rate of temperature change can be expressed by the following equation:

$$\frac{\Delta T}{\Delta t} = \frac{-V_N \bullet V_T / \Delta z}{120} \quad (°\text{C/hr}) \qquad (5)$$

where V, is expressed in kt and the vertical wind shear is expressed in kt per 1000 ft. If the temperature gradient has the same direction as V_N there is cold air advection. Conversely, if the temperature gradient has the opposite direction to V_N there is warm air advection.

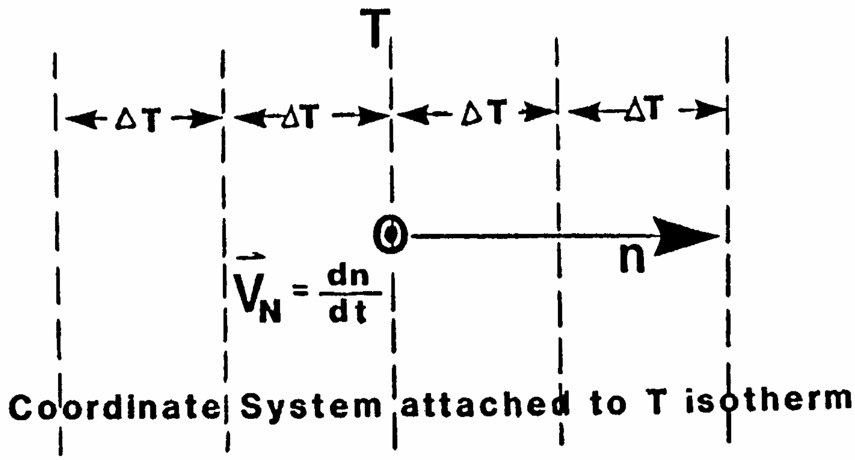

Figure 18.

For instance, in a warm advection situation, if we would find for a 5000 feet thick layer a V_N =30 kts (here V_N is negative since its direction is in the opposite one of the thermal gradient) and a thermal wind vector V_T = 40 kts, the thermal advection would be:

$$\frac{\Delta T}{\Delta t} = \frac{-(-30)\times 40/5(kts/1000\,ft)}{120} = 2\ ^{\circ}\mathrm{C/hr}$$

Horizontal Temperature Gradients

The horizontal spacing of isotherms in frontal zones may be estimated from the geostrophic thermal wind equation in one of the following practical forms:

(a) $$\frac{V_T}{\Delta z}(kt/1000\,ft) = \frac{1}{10}V_{gT}(kt) \qquad (6)$$

where V_{gT} is the fictitious wind found by applying the geostrophic wind scale for isobaric surfaces to isotherms drawn at 5°C intervals and treating them as contours drawn at 200-gp ft intervals. Using this form of the thermal wind equation, the normal spacing of isotherms is expressed as a distance on the geostrophic wind scale.

(b) No. of degrees of latitude between isotherms at 5°C intervals

$$\approx \frac{10}{V_T/\Delta z}(kt/1000\,ft) \qquad (7)$$

Example

The analysis and interpretation of temperature advection for the hodograph in Figure 17 would proceed along the following lines:

For segment CD:

The normal distance between mean isotherms drawn at 5°C intervals in the mixing zone is about one and a quarter degrees of latitude, or 75 nautical miles.

Since V_N is zero, the temperature advection is also zero.

For the segment EF:

The normal distance between 5°C isotherms in the mixing zone is about 100 nautical miles.

It is a cold front moving eastward at around 10 kt. The advective rate of temperature change in the frontal zone is -0.5°C per hour.