ON THE DISTRIBUTION OF TEMPERATURE AND WIND IN THE UPPER WESTERLIES

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ABSTRACT

Systematic studies of certain meridional cross sections carried out at the University of Chicago indicate that the westerlies in the upper troposphere and at the level of the tropopause often show a very narrow zone with extremely strong winds in the region above the frontal layer between the subtropical and the polar air. An example showing the method of analysis and its principal results is given. The stability of the strong westerly current (the "jet stream") in this type of zonal air flow and the corresponding temperature distribution at the level of 200 mb are discussed.

1. Introduction

In most studies of the meridional distribution of temperature and wind (pressure) in the northern hemisphere the discussion has been based upon mean values. However, it is obvious that every study of the three-dimensional temperature, pressure and wind field based upon such data is likely to eliminate features in the atmospheric circulation which are very characteristic of the circulation pattern in singular cases, and therefore, of greatest importance for a real understanding of the whole mechanism.

Recently the Headquarters of U. S. Air Weather Service published several volumes containing daily 500-mb charts for the Northern Hemisphere [9]. These charts will certainly be of much value for the further study of the atmospheric circulation, if publication can be continued for a longer period.

The most striking feature of these charts for the 500-mb level is, probably, that *the middle-latitude belt of westerlies generally appears as a surprisingly narrow band of strong winds, embedded in relatively stagnant air masses to the south and north.* The band of strong westerlies, however, is not always situated at the same latitude. It shows strong latitudinal displacements from day to day and from longitude to longitude and has therefore been described as a "meandering river" [8]. In some regions and at some times this "meandering river" becomes rather indistinct or forms cyclonic or anticyclonic eddies, but in such cases there is always a tendency for a reforming of the strong belt of westerlies later on.

From this picture of the westerlies it follows that the strong concentration of the zonal wind in a narrow zone, so characteristic of the daily charts, almost completely disappears in maps showing the average conditions for longer periods of time. This is due partly to the meridional displacements of the zone of strongest westerlies from day to day and from season to season and partly to the appearance of strong meridional components of the air flow in connection with the formation of the large-scale disturbances characteristic of the middle latitudes.

Also from the 500-mb charts it follows that *the temperature contrast between high and low latitude to a great extent is concentrated in the zone of strongest westerlies and migrates with this zone.* In many cases the 500-mb charts give the impression of a continuous frontal zone running around the hemisphere almost exactly in the belt of strongest wind at this level or a little to the north of it. In other words, there is a very pronounced correlation between horizontal temperature gradient and wind velocity at the level of the 500-mb surface.

A parallel study of the surface maps and the maps for 500 mb makes it obvious that the more or less distinct frontal zones at the 500-mb level, characterized by strong wind, have a clear connection with the surface fronts. It is often not possible to connect minor perturbations in the polar front at the surface level with the band of strong isothermal concentration at the 500-mb level, but the large-scale displacement of the surface fronts can usually be related to similar displacements of the frontal zone at 500 mb. In many cases where it is impossible to define any real front on the surface charts over large areas, there is still a remarkably well-developed frontal zone at the 500-mb level. It can be mentioned that the systematic study of frontal contour charts carried out in the Meteorological Service of Canada has led to similar results [2].

A brief summary of some results achieved in systematic investigations of the characteristic tempera-

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ture and wind field of the westerlies carried out at the University of Chicago during the season 1946–47 will now be given. In a forthcoming paper a more detailed description of the temperature and wind field in a case with almost strictly westerly flow will be presented.

2. Temperature and wind distributions in meridional cross sections

In order to get a detailed picture of the distribution of temperature and wind along a meridional cross section it is necessary to examine cross sections with a relatively dense network of aerological stations extended over such a great distance that the cross section really can give the characteristic distribution of the elements in the whole region between the tropics and the arctic. It is not easy to find many such cross sections in the northern hemisphere.

The network of radiosonde stations in North America is the most suitable for our purpose. The whole investigation, therefore, was based upon these observations. In the future it will be necessary to extend similar investigations to other parts of the northern hemisphere, but there are no indications that the results achieved over North America would not be applicable to the whole hemisphere, with necessary modifications because of the influence of the continents on the general circulation pattern.

The radiosonde stations most suitable for a detailed study of the meridional distribution of temperature and zonal wind in North America are those situated almost along the meridian 80°W. Along this meridian or in its vicinity it is possible to get regular daily radiosonde data from the following stations (international index numbers in parentheses): Havana (030), Miami (202), Tampa (211), Apalachicola (220), Charleston (208), Greensboro (317), Huntington (425), Washington (405), Pittsburgh (520), Buffalo (528), Sault Ste. Marie (734), Moosonee (836), Port Harrison (907), Coral Harbour (915), Arctic Bay (918), and Thule (202) (Greenland).

From the upper-air charts for the whole northern hemisphere, one can see that eastern Canada is the region of a quasistationary upper trough. Therefore, the meridian selected for the detailed study of the temperature and wind field on the average gives the distribution of these meteorological elements in the region of an upper trough. In application of the results on other parts of the northern hemisphere this circumstance must be observed. Over the western parts of the North American continent the situation is not quite so favorable because of the strong influence of the Rocky Mountains on the temperature and wind field.

In a later, extended report the results of systematic studies of the temperature and wind field in the meridional cross section along the meridian 80°W will be discussed in more detail. In particular, the question of the average conditions for longer times (seasons) must be postponed. Here only some special results of interest will be discussed.

As a characteristic example of a vertical cross section along the meridian 80°W the case of 17 January 1947 will be discussed. The surface map and the corresponding 500-mb chart are given in figs. 1 and 2 and the cross section showing the vertical distribution of temperature and zonal (geostrophic) wind in fig. 3.

The polar front, which according to the surface chart was situated just north of Charleston, reaches the 700-mb surface over Greensboro and the 500-mb level over Pittsburgh, and becomes diffuse at the level of 400 mb above Buffalo. The corresponding average slope of the frontal layer is about 1/130. North of the principal front there are other weaker frontal layers separating the very cold "arctic" or polar-continental air masses from the somewhat warmer modified air masses of polar origin farther south.

In the analysis of this cross section the tropopause was not regarded as a continuous surface of discontinuity, but as a boundary layer with different surfaces of discontinuity characterized by quasi-isentropy [1, 4]. Over the southern part of the cross section one can observe the real tropical or equatorial tropopause. Its exact position, however, is difficult to determine because of lack of observations. In the region above the polar-front layer there is a much lower tropopause with an approximate potential temperature of 340A. This tropopause is very well marked in its central parts and shows its maximum elevation at about latitude 35°N. *Both to the south and to the north, this extratropical tropopause sinks.* This is an important fact which will be referred to later on. It might be emphasized that this type of tropopause appears to be rather common over the U. S., in situations with a well-developed polar front in upper levels.[2]

In the region where the frontal layer reaches the tropopause, it is very difficult to recognize one distinct tropopause surface. That is the region where, according to the conclusions derived from investigations of European cyclones, the tropopause shows a very pronounced multiple character [1]. With the aid of a much denser network of radiosonde observations it would probably have been possible to recognize different surfaces of discontinuous change in the lapse rate just in this region of transition to the polar atmosphere.

North of the frontal zone one can notice a distinct polar tropopause at a height of about 9 km with a weak lowering just above the low-pressure center at Hudson Bay. In this region there is a tendency for the formation of a lower tropopause, at approximately 7

[2] The analysis in fig. 3 differs in some details from the same cross section published earlier [8].

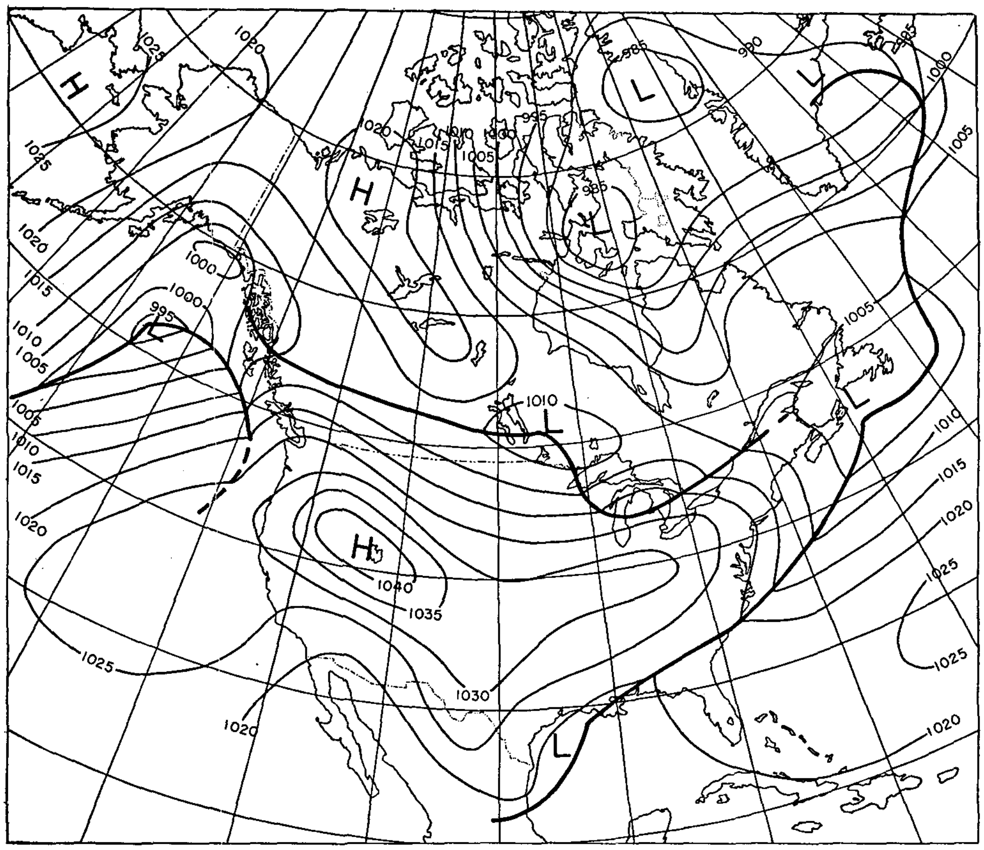

FIG. 1. Surface map 0630 GCT 17 January 1947.

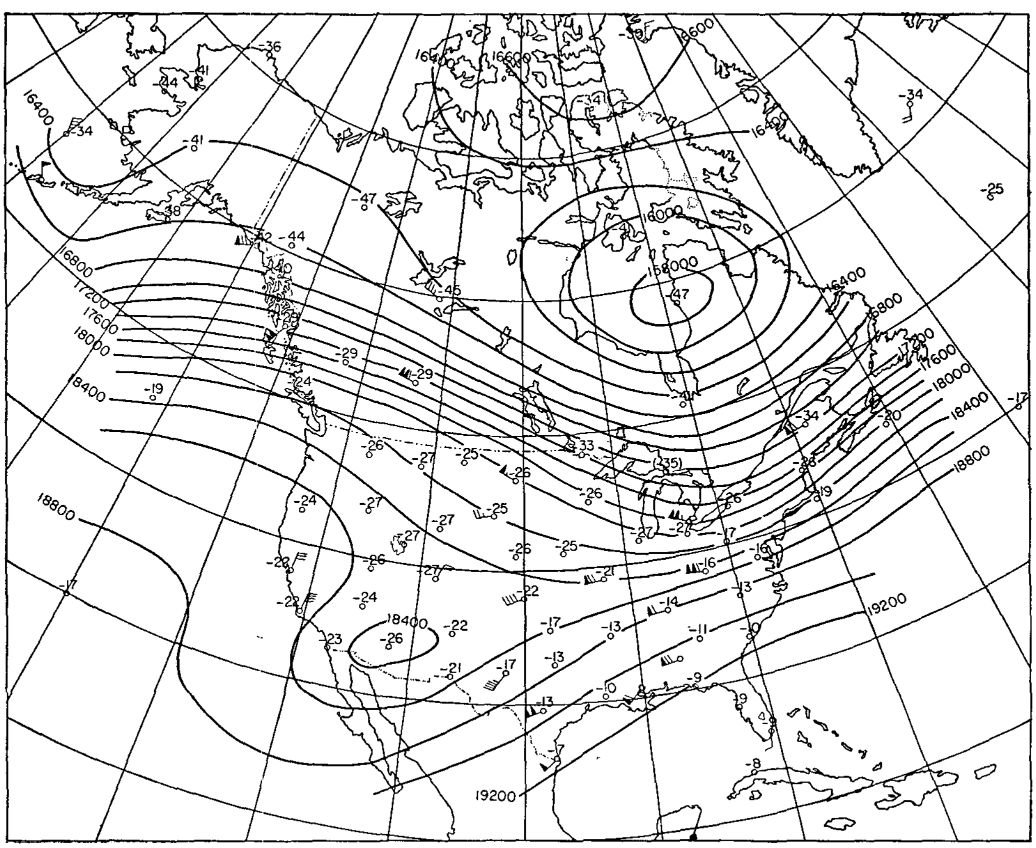

FIG. 2. Contours (ft) and temperatures (C) of the 500-mb surface 0300 GCT 17 January 1947.

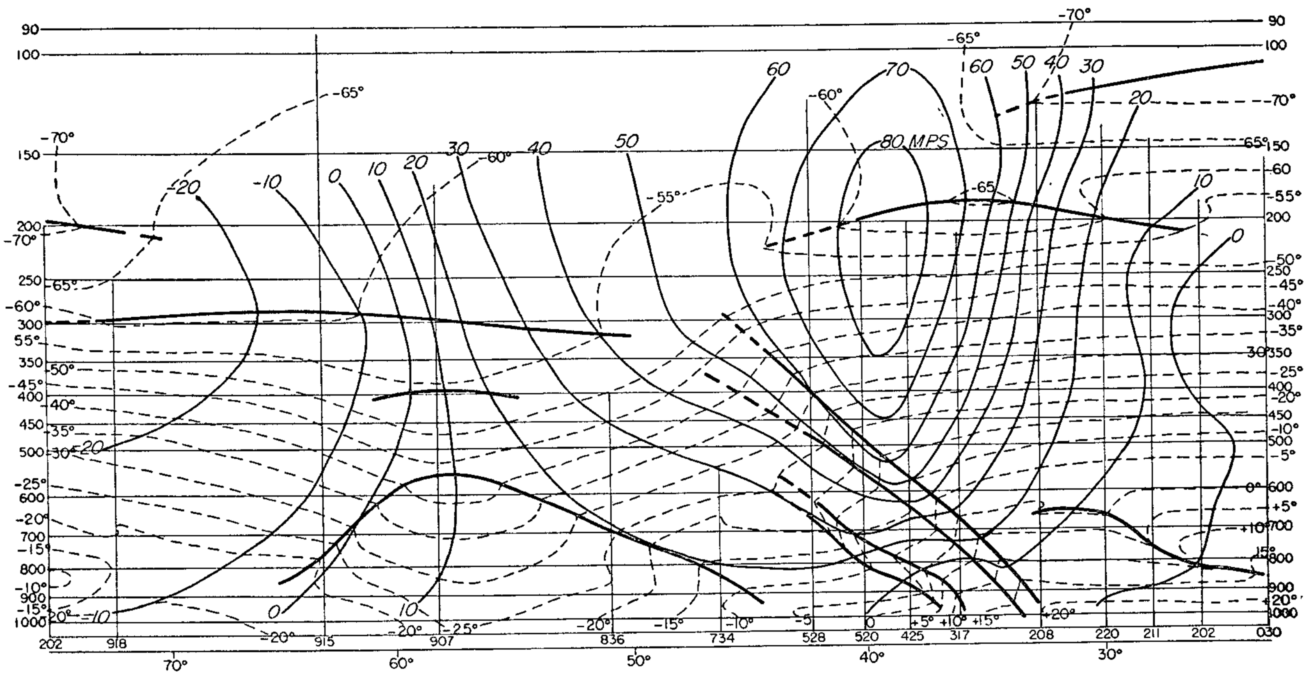

FIG. 3. Vertical cross section approximately along the meridian 80° W from Havana in the south to Thule (Greenland) in the north, Vertical lines indicate the soundings used in the diagram with the international station numbers below. Frontal boundaries, inversions or tropopause surfaces are indicated by thick, solid lines when they are distinct, and by thick, dashed lines when not distinct. Thin. solid lines indicate geostrophic wind velocity (meters per second) perpendicular to the cross section (zonal wind), dashed lines isotherms (C).

km. Farther to the north the temperature of the stratosphere decreases and becomes remarkably low over the arctic regions. Thus, the stations Arctic Bay and Thule indicate temperatures lower than −70C at 12 km. It might be pointed out that *almost all observations in this region during the midwinter season 1946–47 indicate that the arctic stratosphere during the time of no insolation is very cold* and that *the highest stratospheric temperature in wintertime can be found at latitudes 50°–60°N.* Earlier baloonsonde observations in northern Scandinavia indicate the same cold stratospheric air in the arctic regions north of Europe [5].

The meridional cross section from Florida to Alaska drawn and discussed by Willett [11] does not show very much of this low arctic temperature. An explanation of this obvious discrepancy can probably be found in the fact that Willett used radiosonde observations from Alaska where the average temperature values are strongly influenced by advection of air masses from the south in connection with the formation of an upper-air ridge in these regions. Furthermore, Willett used data for January and February in his cross section. In February, however, the stratosphere temperature over the Arctic already begins to increase.

The most interesting feature of the cross section for 17 January (fig. 3) is the geostrophic wind distribution calculated from the slope of the isobaric surfaces. In every level from about 600 mb to the tropopause the strongest wind appears at essentially the same latitude. From fig. 3, one gets the impression of *the existence, at the tropopause level, of a very strong and rather narrow westerly "river," the velocity of which gradually decreases downward and which becomes diffuse at the level of the frontal layer.* The horizontal wind shear to the south of the velocity maximum is very strong. At the level of the strongest wind there is, between latitudes 32°N and 37°N, a shear of the order of magnitude of *10 m sec^{-1} per degree latitude.* As can be seen from fig. 2, the curvature of the wind trajectories in this region of the cross section is rather negligible. The geostrophic wind therefore, computed from the slope of the isobaric surfaces, approximately corresponds to the real wind.[3]

The cross section in fig. 3 gives the characteristic temperature and wind distribution along a meridian for the case of a strong westerly flow and a well-developed polar front. In other situations the picture naturally differs in details, but some characteristic features can usually be observed. The principal results achieved through systematic studies of meridional cross sections can be summarized as follows:

1. The contrast in temperature between the equatorial and polar regions is, in the mean troposphere, to a large extent concentrated in a rather narrow band, which also is the zone of strongest westerlies.

2. The solenoid field in the region of the strongest westerlies is usually composed of two different parts: the solenoid field of the sloping frontal layer character-

[3] The influence of curvature will be discussed in section 3.

ized by great vertical stability (weak lapse rate or inversion), and the solenoid field of the warm air mass above the front characterized by weak vertical stability (strong lapse rate).

3. The strongest west wind can regularly be observed vertically above the latitude where the polar-front layer reaches the height of 600–500 mb. Below this level the horizontal wind shear is rather weak and above this level the air moves as a narrow "river" with the strongest speed at the tropopause level. This zone of strongest wind can be characterized as a "*jet stream*" embedded in the upper westerlies.

4. Southward from the center of the "jet stream" there is a zone several degrees of latitude in width, with a very strong anticyclonic wind shear, at the rate of 10 m sec^{-1} per degree latitude, or even more.

5. In many cases there is a region with lowest temperature in the troposphere and highest temperature in the stratosphere at latitudes 45°–65°N. To the north of this region one can sometimes observe easterly winds in the whole troposphere and in the lower stratosphere; in other cases the westerly winds here are weak.

6. The arctic stratosphere is, during midwinter (at least in December and January), characterized by very low temperature (around −60 to −70C or less). At the level of 200 mb the meridional cross sections therefore give the picture of a cold arctic stratosphere, a ring with higher temperature at latitudes 45°–65°N, a ring with low temperature south of the "jet stream" and then increasing temperature southward to the equator.

3. Stability of the upper westerlies

Some of the results of the systematic analysis of meridional cross sections, especially during the cold season, are illustrated schematically in fig. 4, showing the characteristic distribution of wind in the vicinity of a frontal surface. Two questions of special interest connected with this wind distribution have to be discussed more in detail. As follows from the cross sections, the horizontal wind shear is remarkably strong in the upper troposphere and in the region of the tropopause. To the south of the latitude of maximum wind velocity the wind shear is so strongly anticyclonic that the question of the dynamic stability of the zonal flow arises. Through recent work by Solberg [7], Høiland [3], Van Mieghem [10], and others, the question of the stability of a circular vortex has been introduced in meteorology. The extremely strong zonal flow in our cross sections has to be examined from this point of view. The other question of interest is the obvious difference between the picture given here and the scheme for the wind distribution in the vicinity of a frontal surface, well known in meteorological literature as the scheme of Margules.

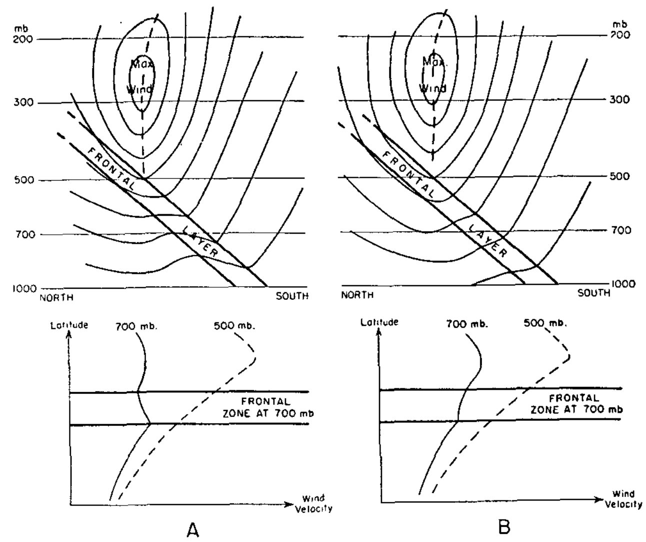

FIG. 4. Schematic picture showing the wind distribution in the vicinity of a frontal layer for a case with strong concentration of solenoids in the frontal layer (A) and for a case with relatively weak concentration (B). The lower part gives the wind velocity at 700 mb (solid curves) and 500 mb (dashed curves); in case A there is cyclonic shear in the frontal zone at the level of the 700-mb surface, in case B anticyclonic shear.

The absolute angular momentum of the westerlies, Ω, is composed of the angular momentum of the earth at a given latitude, Ω_e, and the relative angular momentum Ω_u of the westerly flow. Thus

$$\Omega = \Omega_e + \Omega_u, \tag{1}$$

where

$$\Omega_e = \omega R^2 \cos^2 \phi, \quad \Omega_u = Ru \cos \phi. \tag{2}$$

In these formulae ω is the angular velocity of the earth, R the radius of the earth, ϕ the latitude and u the speed of the west wind. In the case of stability the absolute angular momentum must decrease with increasing latitude. From (1) and (2) the relationship between horizontal wind shear, latitude, and velocity in the case of constant absolute angular momentum is found to be

$$\frac{\partial u}{\partial y} = 2\omega \sin \phi + \frac{u}{R} \tan \phi, \tag{3}$$

where y is the linear coordinate for the latitude ($y = R\phi$). If the shear is greater than the right-hand expression then the motion is unstable.

It might be pointed out that the second term in (3) in most cases is small and therefore can be neglected in the first approximation. The first term is the Coriolis parameter which in middle latitudes has an approximate value of 10^{-4} sec^{-1} corresponding to a shear of 10 m sec^{-1} $(100 \text{ km})^{-1}$. For a wind velocity of 64 m sec^{-1} the second term is 1 m sec^{-1} (100 km^{-1}) at 45°N. Thus for very strong winds the second term is not quite negligible.

The wind in our cross sections was computed from the slope of the isobaric surfaces and represents therefore the geostrophic velocity. Considering the curvature of the trajectories the gradient wind can be calculated and introduced in formula (3). This formula, however, expresses the condition that the absolute vorticity vanishes in the case of transition to instability. If we express the absolute vorticity ζ_a in terms of the shear and curvature of the zonal current,

$$\zeta_a = -\frac{\partial u}{\partial y} + \frac{u}{r} + f.$$

Here r is the radius of curvature of the trajectory and $f = 2\omega \sin\phi$ the Coriolis parameter. In the case of zero vorticity

$$\frac{\partial u}{\partial y} = \frac{u}{r} + f = \frac{u_g}{u} f \tag{4}$$

where u_g is the geostrophic wind. Since the radius of curvature of a parallel of latitude is $r = R/\tan\phi$, we get from (4) the following relation between zonal wind and geostrophic wind velocity:

$$u_g = u + \frac{u^2}{fR}\tan\phi. \tag{5}$$

The second term on the right is smaller than the first, but in cases with very strong wind it cannot be neglected. Assuming a wind velocity of 80 m sec^{-1} at latitude 45°N, the value of the second term is approximately 10 m sec^{-1}, which gives a geostrophic wind of 90 m sec^{-1}.

It is convenient to have a formula which expresses (3) in terms of the geostrophic shear. With use of polar coordinates (3) can be written

$$\frac{\partial}{\partial\phi}(u\cos\phi) = 2\omega R \sin\phi\cos\phi. \tag{6}$$

Differentiating the expression for geostrophic wind in (5) we get, using (6),

$$\omega R\left(\cos\phi\frac{\partial u_g}{\partial\phi} - u_g\sin\phi\right) = 2\omega^2 R^2 \sin\phi\cos\phi + u\frac{\partial u}{\partial\phi}.$$

If one observes that according to (4)

$$u\frac{\partial u}{\partial\phi} = 2\omega R u_g \sin\phi,$$

the expression for the geostrophic shear can be written in the form

$$\frac{\partial u_g}{\partial\phi} = 2\omega R\sin\phi + 3u_g\tan\phi.$$

With use of linear coordinates as in (3), this formula can be written

$$\frac{\partial u_g}{\partial y} = 2\omega\sin\phi + \frac{3u_g}{R}\tan\phi. \tag{7}$$

Equation (7) expresses the relationship between geostrophic wind shear, geostrophic wind, and latitude in the case of constant absolute angular momentum.

The expressions in (3) and (7) show that the influence of curvature in the case of strong zonal flow is considerable. Thus, it is important to use formula (7) when the winds are determined from the slope of the isobaric surfaces or from the pressure field. For a geostrophic wind of 80 m sec^{-1} and a latitude of 45°N the second term in (7) is about 3.7 m sec^{-1} (100 km)$^{-1}$, which has to be compared with the second term in (3).

Equations (3), (4), and (7) determine the critical wind shear or geostrophic wind shear corresponding to the case of dynamic instability. In an atmosphere with static stability (increasing potential temperature with height) the wind shear must be computed along isentropic surfaces, which generally will result in a greater shear than if computed in a horizontal surface. Because of the difficulty of determining the geostrophic wind for the upper atmosphere with the accuracy required by the above formulae, it is extremely difficult to decide whether meridional cross sections in actual cases really show a zone of instability. In a later paper this question will be studied more in detail.

The wind distribution in the vicinity of the front, according to figs. 3 and 4, deviates widely from the scheme well known in meteorological literature, as already pointed out. In meteorological textbooks a front is characterized by cyclonic wind shear separating two relatively broad air currents. In our cross sections the horizontal shear in the upper, warm air mass is very considerable, and the anticyclonic shear to the south of the zone of maximum wind velocity is superimposed on the pressure field of the frontal layer in such a fashion that in many cases no cyclonic wind shear at the front below 600 mb can be observed (case B in fig. 4).

Let u_2 and u_1 denote the wind velocity perpendicular to the cross section at two different heights on the upper boundary of a frontal layer (fig. 5). Since the discontinuity in question is of the *first order* in temperature, there will be no discontinuity in u and hence

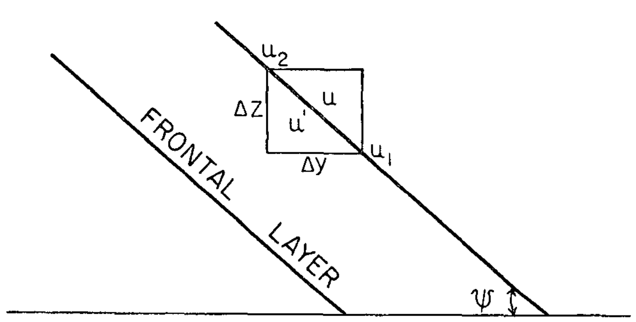

FIG. 5. Schematic diagram for the calculation of wind shear in the vicinity of a sloping frontal layer.

we may write

$$\frac{\partial u}{\partial y}\Delta y + \frac{\partial u}{\partial z}\Delta z = u_2 - u_1 = \frac{\partial u'}{\partial y}\Delta y + \frac{\partial u'}{\partial z}\Delta z, \quad (8)$$

where the primed quantities refer to properties of the frontal zone. Writing $\tan\psi = \Delta z/\Delta y$ for the slope of the front, we see from (8) that

$$\frac{\partial u'}{\partial y} = \frac{\partial u}{\partial y} + \tan\psi\left(\frac{\partial u}{\partial z} - \frac{\partial u'}{\partial z}\right). \quad (9)$$

Using the thermal wind equation in the approximate form

$$\frac{\partial u}{\partial z} = -\frac{g}{fT}\frac{\partial T}{\partial y},$$

it follows from the continuity of temperature ($T = T'$) that (9) may be written

$$\frac{\partial u'}{\partial y} = \frac{\partial u}{\partial y} + \frac{g\tan\psi}{fT}\left(\frac{\partial T'}{\partial y} - \frac{\partial T}{\partial y}\right). \quad (10)$$

The shear $\partial u'/\partial y$ becomes positive (anticyclonic shear) if the expression on the right-hand side is greater than zero. Thus, the inequalities

$$\frac{g\tan\psi}{fT}\left(\frac{\partial T}{\partial y} - \frac{\partial T'}{\partial y}\right) \begin{matrix} > \\ < \end{matrix} \frac{\partial u}{\partial y} \begin{matrix} \text{(cyclonic shear)} \\ \text{(anticyclonic shear)} \end{matrix}$$

express the conditions for cyclonic or anticyclonic shear.

If the anticyclonic shear in the warm air mass is greater than the term arising from the discontinuity of temperature gradient, the west wind increases toward the north in the frontal zone. Extending the above reasoning to the entire frontal layer one therefore can conclude that it is possible to have fronts with anticyclonic shear. In the case of strong concentration of solenoids in the frontal layer, the cold-air current usually is weaker than the warm-air current.

Equation (10) gives the condition for disappearance of the horizontal wind shear at the front. It might be interesting to consider the conditions for which this occurs. Assuming for the inclination of the frontal surface the value 10^{-2}, for f the value 10^{-4} sec^{-1}, and for T the value 250A, we find $g\tan\psi/fT \approx 4$ m sec^{-1} per degree. If the anticyclonic shear in the warm air is more than four times the difference in horizontal temperature gradient between the frontal zone and the warm air just at the front, the front will be characterized not by cyclonic shear, but by anticyclonic shear. Assuming that the maximum anticyclonic shear in the warm air corresponds to constant absolute angular momentum according to formula (3), we can conclude that the anticyclonic shear is not likely to ex-

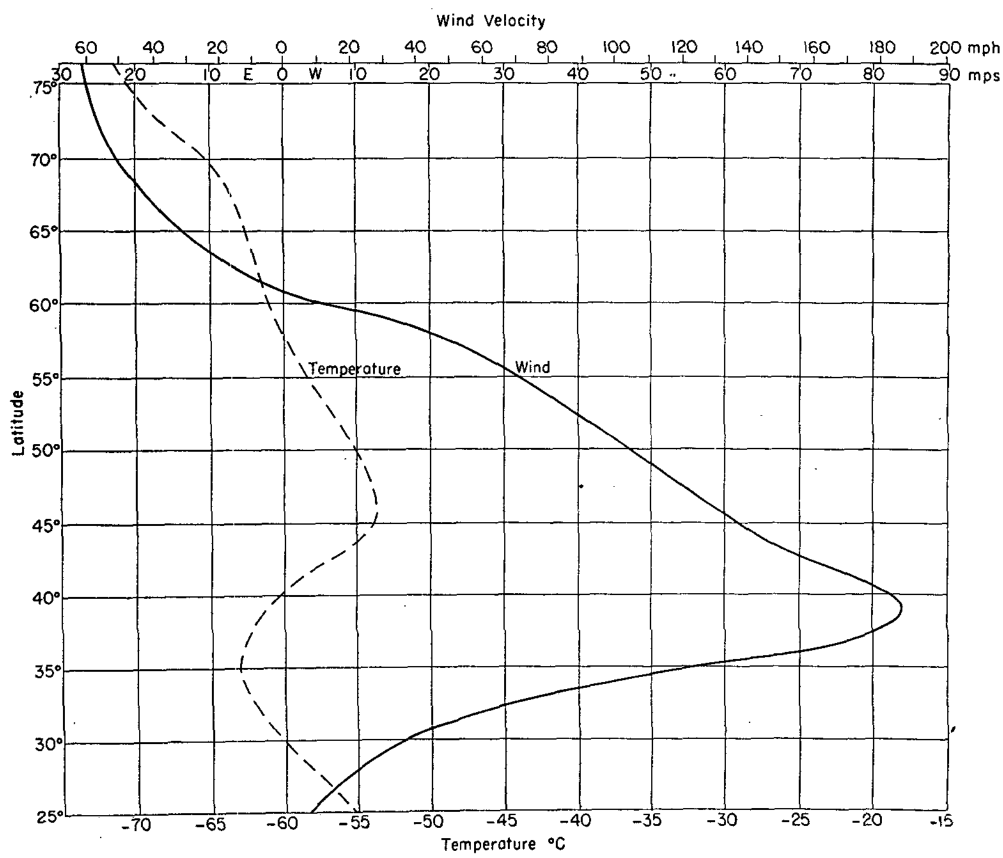

FIG. 6. Distribution of temperature (dashed curve) and geostrophic zonal wind (solid curve) at the 200-mb level along the meridian 80° W for 17 January 1947.

ceed the value 10^{-4} sec^{-1}. The value of $\partial T/\partial y - \partial T'/\partial y$ is then 2.5C (100 km)$^{-1}$, for the case of vanishing shear in the frontal zone. Corresponding to a horizontal temperature gradient of 1C (100 km)$^{-1}$ in the warm air, the temperature gradient in the frontal zone must be greater than 3.5C (100 km)$^{-1}$ to give anticyclonic shear in the frontal zone.

In fig. 4 the horizontal distribution of the wind at the level of 700 mb is represented schematically for the case of cyclonic shear (A) and anticyclonic shear (B). In the first case we assume such a strong frontal solenoid field that the strong anticyclonic shear of the upper warm-air current changes into cyclonic shear at 700 mb. In the second case the frontal solenoid field is not strong enough to change the sign of the shear. It might be emphasized that a wind distribution of the type B rather often can be observed at the 700-mb level. Because of the characteristic location of the region of strongest wind in a cross section (see figs. 3 and 4) it is, however, not likely to find frontal zones with anticyclonic wind shear in the upper part of the troposphere (above 600 or 500 mb).

The relationship between temperature and wind distribution discussed here is not limited to cases with zonal wind; it can be applied to fronts of arbitrary orientation.

4. Some characteristics of the meridional temperature field in the upper atmosphere

Fig. 6 shows the wind and temperature distribution at the level of 200 mb according to the cross section for 17 January 1947. The 200-mb surface intersects the tropopause, as it was drawn in fig. 3, at latitudes 30°N and 40°N; in the lower latitude, however, there is an upper tropical tropopause. The most interesting feature is the relationship between the location of the zone of strongest west wind (jet stream) and the zones of relatively low and high temperature observed at the level of 200 mb. The wind maximum (about 85 m sec^{-1}) can be found at latitude 39°N, the temperature minimum (−63C) at latitude 35°N and temperature maximum (−53.5C) at latitude 46°N. Thus, *there is a zone of lowest temperature about 4 degrees of latitude south of the zone of strongest wind and a zone of highest temperature about 7 degrees of latitude north of the same zone.* A similar result was found in most cross sections which showed a strong concentration of the westerlies.

It is probable that this characteristic relationship between the wind and temperature fields is connected with a special kind of cross-stream vertical circulation superimposed on the zonal motion. The result of this vertical circulation on the temperature field can better be studied using potential temperature. A discussion of this problem will appear in a later paper.[4]

According to fig. 6 the highest temperature at the 200-mb surface in the case of 17 January 1947 can be found at latitude 46°N.[5] Farther north the temperature decreases and at latitude 76°N (Thule) a temperature of about −70C was observed. This observation and most cross sections drawn in the period November–January indicate the existence of a dome of very cold stratospheric air over the north pole. It might be pointed out that the same scheme for the temperature distribution at the level of 200 mb (about 12 km) can be found from the average meridional cross section for January published by the writer in 1934 [5]; this cross section showed a cold ring at latitude 40°N, a warm ring at about latitude 53°N, and a very cold region over the arctic.

The influence of any meridional circulation superimposed on the zonal flow can hardly be studied by means of the temperature distribution in the troposphere because of the strong vertical lapse rate of temperature. However, in the lower stratosphere with its pronounced vertical stability the effect of every vertical component of the air motion on the temperature field is very strong. Therefore, the temperature distribution at the level of 200 mb is especially suitable for the study of the vertical wind field.

[4] See also the discussion in [1].

[5] As mentioned before, the highest stratospheric temperature usually can be found at a somewhat higher latitude.

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