Shreveport to Nashville and then curve upward again to Detroit.

In conclusion, the cross section constructed with isotherms of potential temperature and isotherms of condensation temperature has the following advantages over one drawn which uses temperature and specific humidity: first, it has thermodynamic significance because the nearness to saturation of any point can be determined; second, it roughly shows the convective instability of the air as was pointed out on figure 3; third, the slope of many isentropic surfaces is shown; and finally, because fronts follow isentropic surfaces, the position and slope of the fronts which cut the cross section plane can be identified.

The author wishes to express his gratitude to Dr. H. R. Byers for his able guidance and many helpful comments during the preparation of this paper.

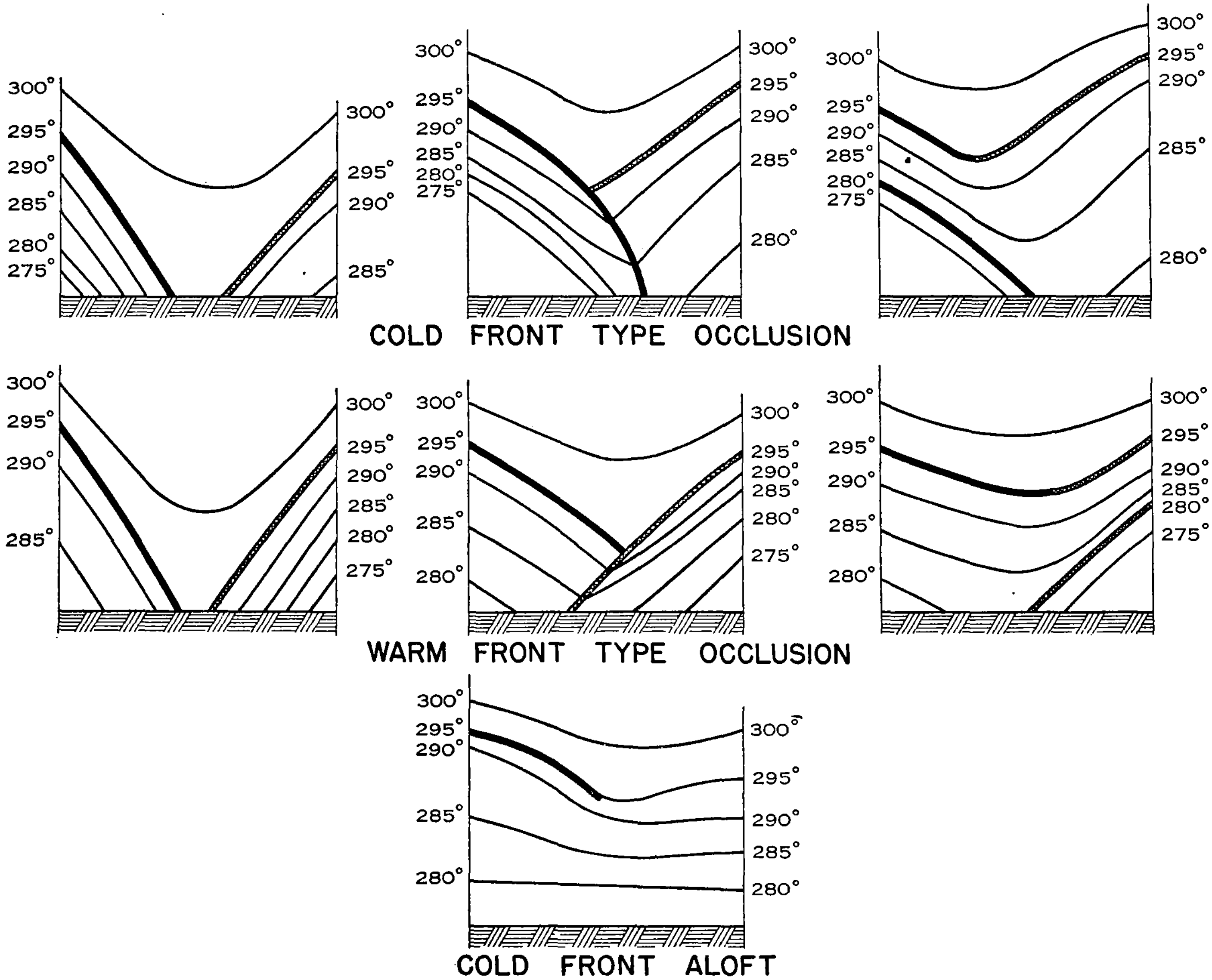

Figure 7.—Cross sections showing possible history of frontal positions on isentropic surfaces in cold- and warm-front-type occlusions and cold front a oft.

A PRACTICAL METHOD FOR COMPUTING WINDS ALOFT FROM PRESSURE AND TEMPERATURE FIELDS

By Edward M. Vernon and E. V. Ashburn

[Weather Bureau, Oakland, Calif., May 1938]

The Weather Bureau's network of pilot-balloon observations supplies, as a rule, all the information on winds aloft required for the operation of aircraft. However, it is not an infrequent occurrence for current wind-aloft data to be missing over a large area, due to the presence of low cloud or other weather conditions which interfere with pilot-balloon observations. It is during such weather conditions, which either prevent or greatly limit pilot-balloon observations, that the airplane pilot is most dependent upon information concerning the winds in the upper air. For this reason it has long been desirable to have a practical method for determining the winds aloft at various altitudes without having to rely entirely upon pilot-balloon data.

Meisinger [1] developed a system for the preparation of free air pressure maps which he hoped would provide a basis for the issuance of wind aloft forecasts. The pressure maps were made possible by the reduction of surface pressures to the 2 and 3 kilometer levels. However, his system, notwithstanding its apparent advantages, has received but little practical application.

Shaw [2] gives a formula for computing the variations of wind with height, while Brunt [3] gives equations resembling those given by Shaw. It appears that neither Shaw nor Brunt regarded the principles as being adaptable to the analysis of current upper wind situations. On this we may quote Shaw:[2] "The several forms of the equation cannot be applied generally to numerical evaluation of special cases of the variation of wind with height in the free air because the individual values of the horizontal gradients * * * and the lapse rates * * * of temperature and pressure are not known * * *"; and Brunt:[3] "These results are not easy to apply to individual cases, * * *." Today, in the United States, with a framework of aero-meteorograph and radio-meteorograph soundings, supplemented by a large mass of free air temperature data being taken at frequent intervals by commercial aircraft operators, it appears that the time has come when practical use can be made of methods similar to those treated theoretically by Shaw and Brunt.[4]

Let us consider for the moment what methods the meteorologist is likely to use today in determining what winds aloft should prevail in an area where pilot-balloon data are missing. To begin with, he will probably draw streamlines on the upper air charts, using the available pilot-balloon data, and then attempt to extend them through the area devoid of data. If this area is small, very good results may be obtained. If it is large, accurate determinations are difficult, frequently impossible. It is possible that this method was first used by Shaw [5], and it has been used at a number of Weather Bureau stations. Recently the advantages to be gained through its application were demonstrated in a paper by Haynes [6], who constructed geostrophic wind scales for drawing streamlines.

If the streamline method fails to give the desired information, or if the data appear to be too meager for accurate results with this method, the meteorologist may measure the pressure gradient on the sea level chart (or the 5,000-foot plane chart in the Plateau Region). From it he may determine the geostrophic wind, which it is assumed should be quite representative of the wind from one to three thousand feet above the base plane, and representative in less degree for higher levels. He may then go so far as to form some rough opinion as to the amount by which the wind direction and speed should change with altitude. This approximation he may base simply on averages of the manner in which the wind changes with height, or he may go to the extent of considering qualitatively the effect of the horizontal distribution of temperature on the change of wind with height. Except for the work done by Haurwitz and Meisinger, it seems that no attempt has been made to reduce to a practical routine this last problem of evaluating the effect of temperature distribution. This paper is, therefore, devoted to the problem of developing a practical method for determining the winds aloft by both qualitative and quantitative consideration of the distribution of pressure and temperature. In order to justify the method, which in its final form will be simple and easily applied, it is necessary first to present the fundamental theory underlying its development.

We begin with the supposition that the wind at any level in the upper air is directly related to the distribution of pressure at that level by the gradient wind equation. Next we accept the well-known equation expressing the pressure—height—temperature relation as being the fundamental concept necessary for the determination of pressures in the free air and therefore of pressure gradients at any upper level.

Writing the hypsometric equation we have:

$$P=P_0 e^{-\left[\frac{g(h-h_0)}{RT_m}\right]} \tag{1}$$

in which: h_0=altitude of lower plane
h=altitude of upper plane
P_0=pressure at altitude h_0
P=pressure at altitude h
e=base of natural logarithms
R=characteristic gas constant for air
T_m=mean virtual temperature of the air between h and h_0
g=mean gravity between h and h_0

The pressure, P, at a fixed upper level, h, is thus a function of the pressure P_0 at a fixed lower level h_0 directly below the point where P is measured, and of the mean virtual temperature, T_m, of the air column vertically between the levels h and h_0 directly below the point where P is measured. Hence we may write:

$$P=f(P_0,\ T_m) \tag{2}$$

For each point represented by the coordinates x, y, in a horizontal surface, there corresponds at any given instant of time a definite set of values P_0 and T_m; that is, at each instant, P is a scalar point function of the coordinates x, y.

The gradient of the function $f(P_0, T_m)=P$ at a point is

$$\nabla P=\mathbf{i}\frac{\partial f}{\partial x}+\mathbf{j}\frac{\partial f}{\partial y} \tag{3}$$

$$=\frac{\partial f}{\partial P_0}\nabla P_0+\frac{\partial f}{\partial T_m}\nabla T_m \tag{4}$$

Here the gradients ∇P, ∇P_0, ∇T_m, are vectors measured positively in the direction of increasing values of P, P_0, and T_m, respectively, oppositely to the direction customarily used in meteorology for the gradient. Equation (4) holds equally well if meteorological gradients are used for the mathematical gradients, since on multiplying both sides of the equation by minus one, the appropriate meteorological gradients may be substituted for $-\nabla P$, $-\nabla P_0$, and $-\nabla T_m$.

The fields of pressure, P and P_0, can be represented by isobars, while the field of mean temperature of the air column, T_m, can be represented by isotherms. If $\boldsymbol{n}_1$ and $\boldsymbol{n}_2$ are unit vectors at a point (x, y) in the fields of P and P_0 measured normal to the isobars in the direction of decreasing values of P and P_0, respectively, and $\boldsymbol{n}_3$ is a unit vector at the point (x, y) in the field of T_m measured

[1] MONTHLY WEATHER REVIEW, Supplement No. 21, 1922; The Preparation and Significance of Free Air Pressure Maps for the Eastern United States.
[2] Manual of Meteorology, vol. IV (1931) pp. 196-7.
[3] Physical and Dynamical Meteorology, pp. 200-201.
[4] B. Haurwitz of the Canadian Meteorological Service has taken the equations presented by Brunt and worked out a method of applying them to actual cases.
[5] Manual of Meteorology, vol. IV (1931), pp. 212-3.
[6] Upper Wind Forecasting; Monthly Weather Review, 66 : 4-6, January 1938.

normal to the isotherms in the direction of decreasing values of T_m, then

$$\left.\begin{aligned} -\nabla P &= \frac{dP}{dn_1}\boldsymbol{n}_1 \\ -\nabla P_0 &= \frac{dP_0}{dn_2}\boldsymbol{n}_2 \\ -\nabla T_m &= \frac{dT_m}{dn_3}\boldsymbol{n}_3 \end{aligned}\right\} \tag{5}$$

whence from (4) and (5)

$$\frac{dP}{dn_1}\boldsymbol{n}_1 = \frac{\partial f}{\partial P_0}\frac{dP_0}{dn_2}\boldsymbol{n}_2 + \frac{\partial f}{\partial T_m}\frac{dT_m}{dn_3}\boldsymbol{n}_3 \tag{6}$$

This is a vector equation in which $\frac{dP}{dn_1}$, $\frac{dP_0}{dn_2}$, and $\frac{dT_m}{dn_3}$ are scalar values of the respective meteorological gradients in question.

The vector $\frac{dP}{dn_1}\boldsymbol{n}_1 \equiv \boldsymbol{V}_1$ on the left side of equation (6) is equal to the vector sum of the two component vectors

$$\frac{\partial f}{\partial P_0}\frac{dP_0}{dn_2}\boldsymbol{n}_2 \equiv \boldsymbol{V}_2 \text{ and } \frac{\partial f}{\partial T_m}\frac{dT_m}{dn_3}\boldsymbol{n}_3 \equiv \boldsymbol{V}_3.$$

The vector sum is of course obtained by finding the resultant of the component vectors $\boldsymbol{V}_2$ and $\boldsymbol{V}_3$ by the parallelogram method. From equation (1) we find the required values of $\frac{\partial f}{\partial P_0}$ and $\frac{\partial f}{\partial T_m}$ and substitute them in the expressions for $\boldsymbol{V}_2$ and $\boldsymbol{V}_3$.

Thus, the meteorological pressure gradient of P at height h is given by the vector sum,

$$\frac{dP}{dn_1}\boldsymbol{n}_1 = \boldsymbol{V}_2 + \boldsymbol{V}_3 \tag{7}$$

where

$$\boldsymbol{V}_2 = \frac{dP_0}{dn_2}\left(e^{-\left[\frac{g(h-h_o)}{RT_m}\right]}\right)\boldsymbol{n}_2 \tag{8}$$

$$\boldsymbol{V}_3 = P_0\left(e^{-\left[\frac{g(h-h_0)}{RT_m}\right]}\right)\left(\frac{g(h-h_o)}{RT_m^2}\right)\frac{dT_m}{dn_3}\boldsymbol{n}_3 \tag{9}$$

The component vectors, $\boldsymbol{V}_2$ and $\boldsymbol{V}_3$, represent components of the pressure gradient at the level, h, and are always directed toward lower values of P_0 and T_m, respectively. They may be converted into geostrophic wind components which will be directed normal to the gradients of P_0 and T_m, i. e., they will be directed cyclonically about lower values of P_0 and T_m respectively.

The equation for the geostrophic wind is written,

$$\frac{dP}{dn} = v2\omega\rho \sin\phi \tag{10}$$

in which dP/dn is the horizontal pressure gradient, v the wind speed, ω the angular velocity of rotation of the earth, ϕ the latitude, and ρ the density of the air.

But by the gas equation,

$$\rho = \frac{P}{RT} \tag{11}$$

in which P and T are the pressure and temperature at the level under consideration. From equation (1), P may be expressed in terms of P_0, whence (11) becomes,

$$\rho = \frac{P_0 e^{\left(\frac{-g(h-h_0)}{RT_m}\right)}}{RT} \tag{12}$$

and (10) becomes,

$$\frac{dP}{dn} = v\ 2\omega \sin\phi\ \frac{P_0 e^{\left(\frac{-g(h-h_0)}{RT_m}\right)}}{RT} \tag{13}$$

Since $\boldsymbol{V}_2$ and $\boldsymbol{V}_3$ are pressure gradients at height h, the right member of equation (13) with v equal to v_2, v_3, respectively, may be substituted for $\boldsymbol{V}_2$ and $\boldsymbol{V}_3$ in equations (8) and (9). Making these substitutions, transposing, and simplifying, we obtain the following equations for the vector components of wind velocity corresponding to the pressure gradient vectors $\boldsymbol{V}_2$ and $\boldsymbol{V}_3$, respectively.

$$v_2 = \frac{dP_0}{dn_2}\cdot\frac{RT}{P_0\,2\omega\sin\phi} \tag{14}$$

$$v_3 = \frac{dT_m}{dn_3}\frac{Tg(h-h_0)}{T_m^2\,2\omega\sin\phi} \tag{15}$$

Thus we see that the wind at any upper level h, is resolved into two orthogonal components v_2 and v_3. v_2 depends largely upon the pressure gradient at the lower level h_0. It has been said of v_2 that it is identical with the geostrophic wind based on the pressure gradient at h_0. This is not strictly true because it depends also upon T, the absolute temperature at height h, which, of course, does not coincide with temperature T_0 prevailing at h_0. We have termed this component the "isobaric" component because its value is determined from the isobars at h_0.

Equations (14) and (15) also follow immediately from equations (8) on page 200 of Brunt, *Physical and Dynamical Meteorology.*

The component, v_3 depends mainly upon the horizontal gradient of mean virtual temperature, $\frac{dT_m}{dn_3}$, and is therefore called the "thermal" component, after Gold.[7]

Thus far all equations have been written for C. G. S. units. For practical purposes it is convenient to write equations (14) and (15) in mixed units:

$$v_2 = \frac{dP_0}{dn_2}\frac{RT}{P_0\,2\omega\sin\phi}\frac{.00000069756}{K} \tag{16}$$

$$v_3 = \frac{dT_m}{dn_3}\frac{T}{T_m^2\,2\omega\sin\phi}\frac{g(h-h_0).00000038754}{K} \tag{17}$$

Here, $R=2.87\times10^6$, $\omega=\frac{2\pi}{86{,}164}$, ϕ, g, h, and h_0 have the same significance as in equation (1) and are in C. G. S. units, while

v_2 and v_3=velocity in miles per hour.
dP_0=Isobaric interval at h_0, in units of 0.05 in mercury.
dT_m=Isotherm (dT_m) interval in units of 5° F.
T=Absolute temperature at height, h.

[7] Shaw. Manual of Meteorology, vol. IV (1931); p. 202.

dn_2 and dn_3=millimeters on map between isobars or isotherms. (If the constant, K, is omitted dn_2 and dn_3 are in statute miles.)

P_0=Pressure at h_0, in hundredths inch mercury, written without decimal.

K=a constant equal to the number of miles represented by 1 mm at latitude ϕ on the particular map projection used.[8]

SCALES FOR MEASURING ISOBARIC AND THERMAL COMPONENTS

Wind scales for measuring the isobaric and thermal components are constructed from equations (16) and (17). They may be used on charts having isobars drawn for P_0 and isotherms drawn for T_m. Figure 1 shows a scale for working from the 5,000-foot plane pressure chart to the 10,000-, the 14,000-, and the 20,000-foot levels. Figure 2 shows a scale for working from the sea level pressure map to the 5,000-, the 10,000-, the 14,000-, and the 20,000-foot levels. In the preparation of the scales, g was set constant at the value obtaining between h and h_0 at latitude 40° N. with only negligible error resulting. The values used for T, T_m, and P_0 were taken from the standard atmosphere and used as constants, for each scale depending only on the appropriate value of h and h_0. Some small errors result from the use of constant values for these elements, whenever T, T_m, and P_0 depart from the standard atmosphere values. However, by using the particular values of T, T_m, and P_0 pertinent to the value of h and h_0, the magnitude of these errors is kept at a minimum. They are consequently quite small and usually negligible. The corrections are calculated as follows:

Corrections for temperature.—Assume that the actual temperature, T', obtaining at height h will depart from the standard atmosphere value of T, used in constructing the scale, in the same amount as the actual value of mean temperature, T'_m, departs from the standard atmosphere value of T_m used in the construction of the scale. Let the departure from the standard atmosphere value of T_m be ΔT_m. Then,

$$T' = T+\Delta T_m$$
$$T'_m = T_m+\Delta T_m$$

It follows from equations (16) and (17) that the percentages of error caused by the departure of the actual temperature from the standard atmosphere value are given by:

(*a*) For the isobaric component,

$$\text{Percentage of error}=\frac{\Delta v_2}{v_2}=100\left(\frac{\Delta T_m}{T}\right)$$

The correction is therefore added for positive values of ΔT_m, i. e., when T'_m exceeds T_m.

(*b*) For the thermal component,

$$\text{Percentage of error}=\frac{\Delta v_3}{v_3}=-100\left[-\frac{T+\Delta T_m}{(T_m+\Delta T_m)^2}+\frac{T}{T_m^2}\right]\frac{T_m^2}{T}$$

The percent of error can now be calculated, since T and T_m are the standard atmosphere values used in constructing the scales, while ΔT_m can be taken from the chart. Appropriate corrections for ΔT_m have been calculated and entered on the scales. It will be noted that they are quite small and that they will usually be negligible.

[8] W. R. Gregg and I. R. Tannehill, International Standard Projections for Meteorological Charts, MONTHLY WEATHER REVIEW, 65: 411-415, December 1937.

Correction for pressure.—The only pressure used in constructing the scales was P_0, which is subject to direct observation and for which no substitution is therefore necessary. Let the actual pressure, P'_0, depart by the amount, ΔP_0, from the standard atmosphere value of P_0 used in constructing the scale. Then,

$$P'_0=P_0+\Delta P_0$$

It follows from equations (16) and (17) that the percentage of error caused by the departure of pressure from the standard atmosphere value is given by:

(*a*) For the isobaric component,

$$\text{Percentage of error}=-100\left(\frac{\Delta P_0}{P_0}\right)$$

Corrections for ΔP_0 have been calculated and entered on the scale for the isobaric component, with explicit directions for their application. It will be observed that they are small and usually negligible. There is no correction for ΔP_0 on the scale for the thermal component.

DESCRIPTION AND USE OF THE WIND SCALES

The scales are constructed by plotting curves for latitudes 30°, 40°, and 50°, based on coordinates of wind velocity in miles per hour and distances, dn_2 and dn_3, between successive isobars, or isotherms, in millimeters, as computed by equations (16) and (17).

Scale for thermal component.—The upper half of figure 1 is a scale for measuring the velocity of the thermal component when the 5,000-foot plane is used as a base, while the upper half of figure 2 is the thermal component scale for use with the sea-level chart as a base. The distances between successive 5° F. isotherms of T_m are laid off vertically as ordinates, positively upward from the x-axis which is marked "zero line" on the figure. From the synoptic chart, the distance between isotherms may be measured with a ruler, or simply marked off on the edge of a strip of paper; then, on figure 1, the point which lies at that distance from the zero line, and on the appropriate latitude curve, is readily located, and vertically above it on the appropriate velocity scale the velocity of the thermal component is read off. A small correction for departure of T_m from the standard value may be applied by referring to the corrections noted on the diagram. This velocity component should be entered on the chart as a vector drawn to any convenient scale, and placed on the synoptic chart at the point under consideration.

The vector for the thermal component, thus determined, must *always* be parallel to the isotherm passing through the point under consideration, and must be directed *cyclonically* about lower values of T_m, i. e., must be drawn so that lower temperatures are to the left of an observer with back to the wind.

It is important to understand that if the velocity of the thermal component for the 10,000-foot plane is desired, the isotherms used with the scale must represent the horizontal distribution of mean temperature applying between the base plane and the 10,000-foot plane. Likewise, if the velocity of the thermal component on the 14,000-foot plane is desired, the isotherms must be drawn for mean temperatures applying between the base plane and the 14,000-foot plane, etc.

Scale for isobaric component.—Scales for determining the velocity of the isobaric component are shown in the lower half of figures 1 and 2, the former for use with the base chart at 5,000 feet and latter with the base chart at sea level. The isobaric interval is 0.05 inch of mercury

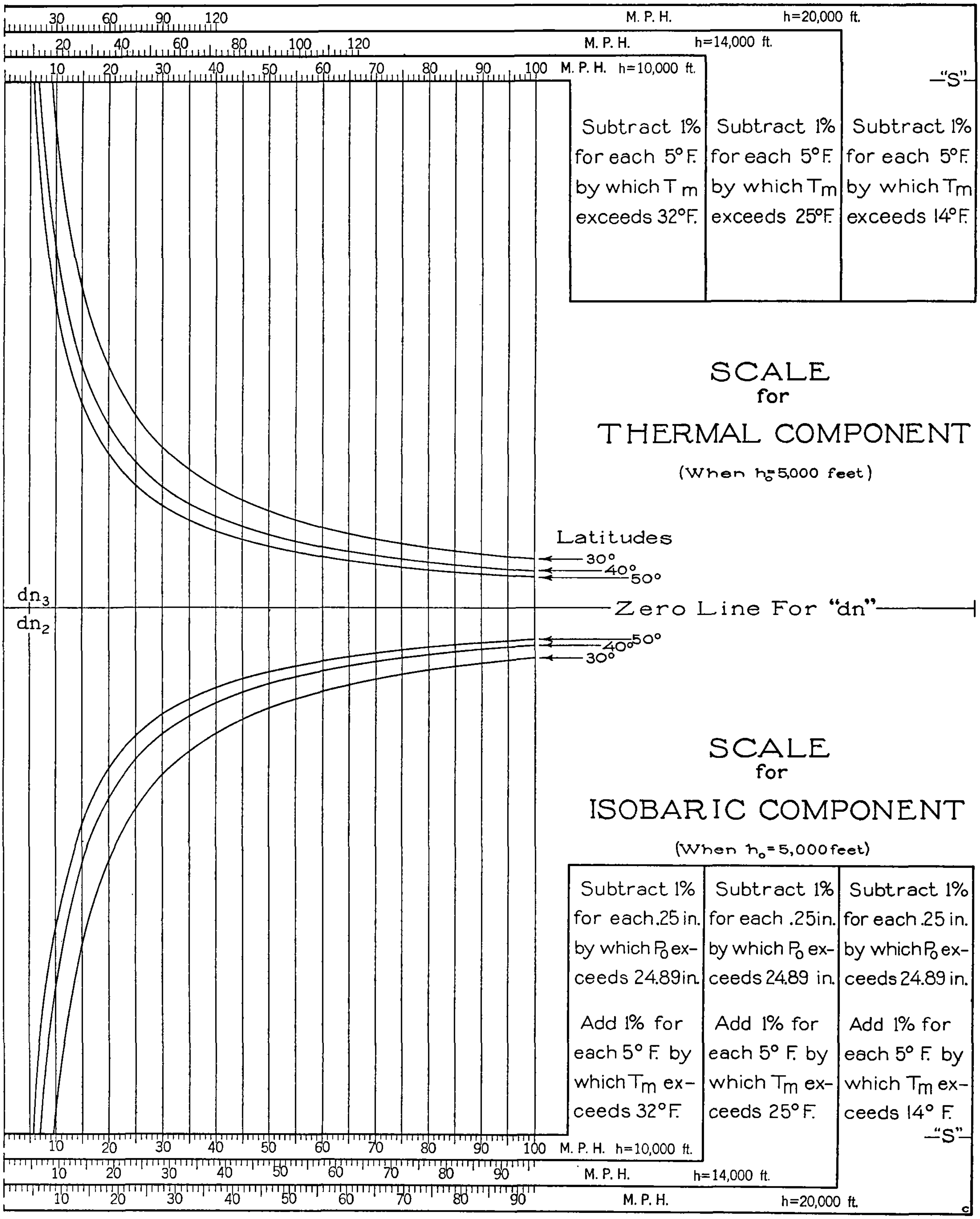
M. P. H. h=20,000 ft.
M. P. H. h=14,000 ft.
M. P. H. h=10,000 ft.
"S"
Subtract 1% for each 5° F. by which T_m exceeds 32°F.
Subtract 1% for each 5° F. by which T_m exceeds 25°F.
Subtract 1% for each 5° F. by which T_m exceeds 14°F.
SCALE for THERMAL COMPONENT
(When h_o=5,000 feet)
Latitudes
30°
40°
50°
dn_3
dn_2
Zero Line For "dn"
50°
40°
30°
SCALE for ISOBARIC COMPONENT
(When h_o=5,000 feet)
Subtract 1% for each .25 in. by which P_0 exceeds 24.89 in.
Add 1% for each 5° F. by which T_m exceeds 32° F.
Subtract 1% for each .25 in. by which P_0 exceeds 24.89 in.
Add 1% for each 5° F. by which T_m exceeds 25° F.
Subtract 1% for each .25 in. by which P_0 exceeds 24.89 in.
Add 1% for each 5° F. by which T_m exceeds 14° F.
"S"
M. P. H. h=10,000 ft.
M. P. H. h=14,000 ft.
M. P. H. h=20,000 ft.

FIGURE 1.

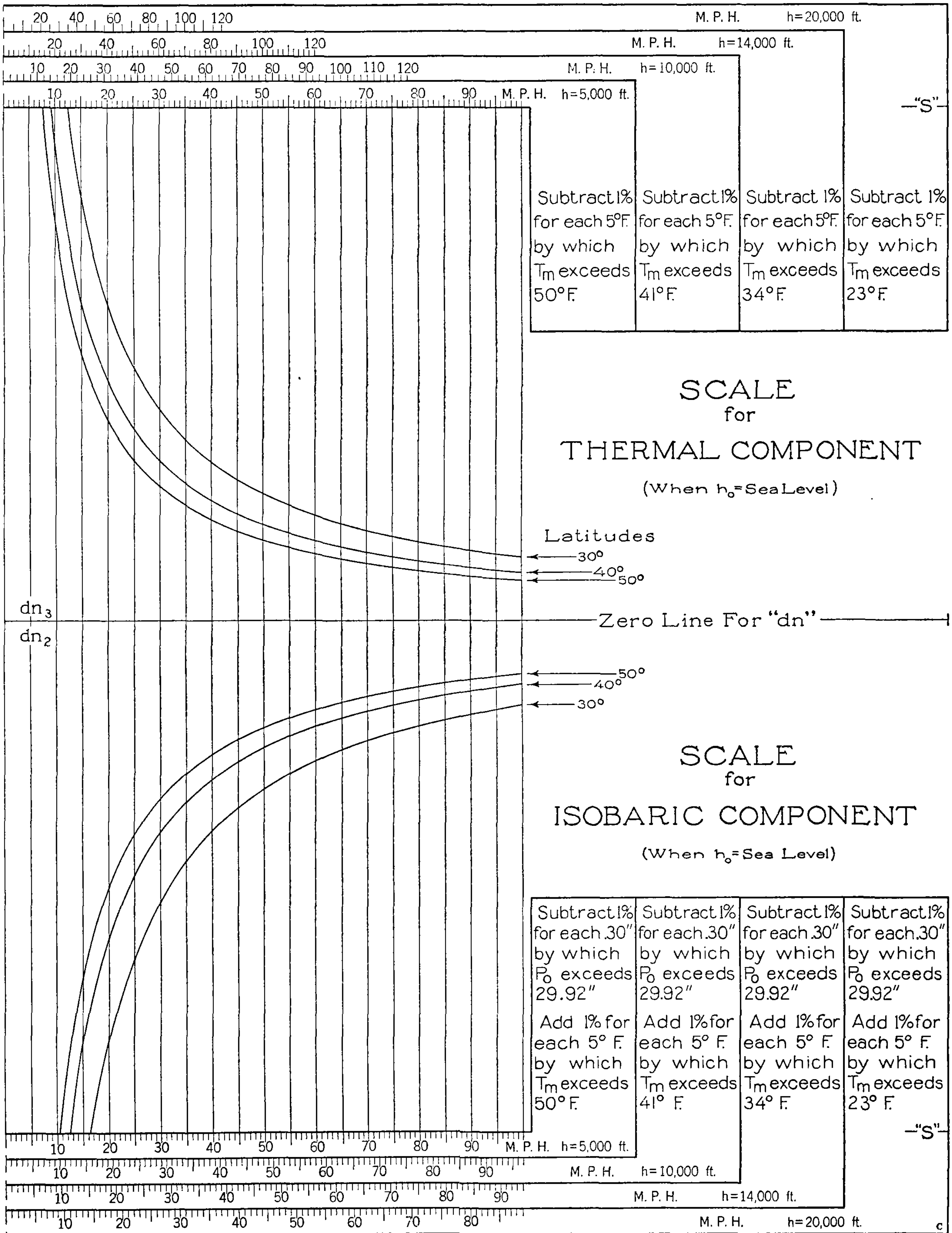
M. P. H. h=20,000 ft.
M. P. H. h=14,000 ft.
M. P. H. h=10,000 ft.
M. P. H. h=5,000 ft.
"S"
Subtract 1% for each 5°F by which Tm exceeds 50°F
Subtract 1% for each 5°F by which Tm exceeds 41°F
Subtract 1% for each 5°F by which Tm exceeds 34°F
Subtract 1% for each 5°F by which Tm exceeds 23°F
SCALE for THERMAL COMPONENT
(When ho=Sea Level)
Latitudes
30°
40°
50°
dn3
dn2
Zero Line For "dn"
50°
40°
30°
SCALE for ISOBARIC COMPONENT
(When ho=Sea Level)
Subtract 1% for each .30" by which Po exceeds 29.92"
Add 1% for each 5° F by which Tm exceeds 50° F
Subtract 1% for each .30" by which Po exceeds 29.92"
Add 1% for each 5° F by which Tm exceeds 41° F
Subtract 1% for each .30" by which Po exceeds 29.92"
Add 1% for each 5° F by which Tm exceeds 34° F
Subtract 1% for each .30" by which Po exceeds 29.92"
Add 1% for each 5° F by which Tm exceeds 23° F
"S"
M. P. H. h=5,000 ft.
M. P. H. h=10,000 ft.
M. P. H. h=14,000 ft.
M. P. H. h=20,000 ft.

FIGURE 2.

on the 5,000-foot chart and 0.10 inch on the sea-level chart. The distances between successive isobars are measured positively downward from the x-axis, marked "zero line" on the sale. If necessary, the corrections for P_0 and T_m noted on the scale are applied.

A vector for the isobaric component, drawn to the same scale as that used in drawing the thermal component, should be drawn on the chart with the tail of the isobaric vector starting at the head of the thermal vector. The isobaric vector must always be parallel to the isobar passing through the point under consideration, and must be directed cyclonically about lower values of P_0.

After the two vectors have been drawn as described, a resultant vector drawn from the tail of the thermal vector to the head of the isobaric vector will give the direction and speed of the wind at the point and altitude under consideration.

The wind scales reproduced in figures 1 and 2 are for use on a conformal conic map projection of the scale 1: 5,000,000 at latitudes 30° and 60°. For use on other conformal conic projections having the same standard parallels, the scales should be reduced photographically so that the mark, "S", appearing at the right edge of each scale is the following distances from the "zero line":

Scale of map projection	Distance "S" should be from the "zero line"
	mm
1:5,000,000	100
1:7,500,000	66.7
1:8,300,000	60.2
1:10,000,000	50
1:15,000,000	33.3

EXAMPLE

For the purpose of giving a practical example of the working of this method, the situation prevailing at 7:30 a. m. May 2, 1938, has been selected. A disturbance was centered in western Utah. Extensive low clouds and precipitation interfered with pilot balloon observations to the extent that no upper wind data were available for the 14,000-foot level from any station to the west or southwest of Rock Springs, Wyoming, and Boise, Idaho. The entire Oakland airway district, extending from Oakland to Medford, Salt Lake City, and Burbank, was without upper wind data for the 14,000-foot level.

The wind for the 14,000-foot plane was calculated for a few points by means of the method just described. Figure 3 shows the chart used for this purpose. The isobars for the 5,000-foot plane are in solid black lines, drawn for intervals of 0.05 inch. Isotherms for temperatures prevailing on the 10,000-foot plane are in broken black lines, drawn for every 5° F. They are assumed to represent approximately the horizontal distribution of mean virtual temperature prevailing between the 5,000- and the 14,000-foot levels. It is far more important to have a correct representation of the horizontal distribution of T_m than to have the exact value of T_m at any one point, since the absolute value of T_m has only a minor effect on the magnitude of the thermal component. The 4 a. m. airplane observations furnished the data for approximate spacing of the isotherms, but it was necessary to resort to temperatures reported by air line pilots in order to obtain the proper spacing of the isotherms between Salt Lake City and Oakland and between Oakland and Portland.

In using air line temperature reports it is not generally advisable to use the actual values of temperature reported, but rather to use the horizontal gradient as shown by a single plane flying at a more or less fixed altitude over a rather long route. For instance, in this example, we already had absolute values of temperature of 25° over Oakland and 29° over Salt Lake City, determined by the airplane meteorograph observations. Analysis of air line temperatures, synchronizing roughly with the airplane observations and available for this analysis, showed that it was 4° colder over Elko than over Salt Lake City while over Medford the temperature was about 6° lower than that over Oakland.

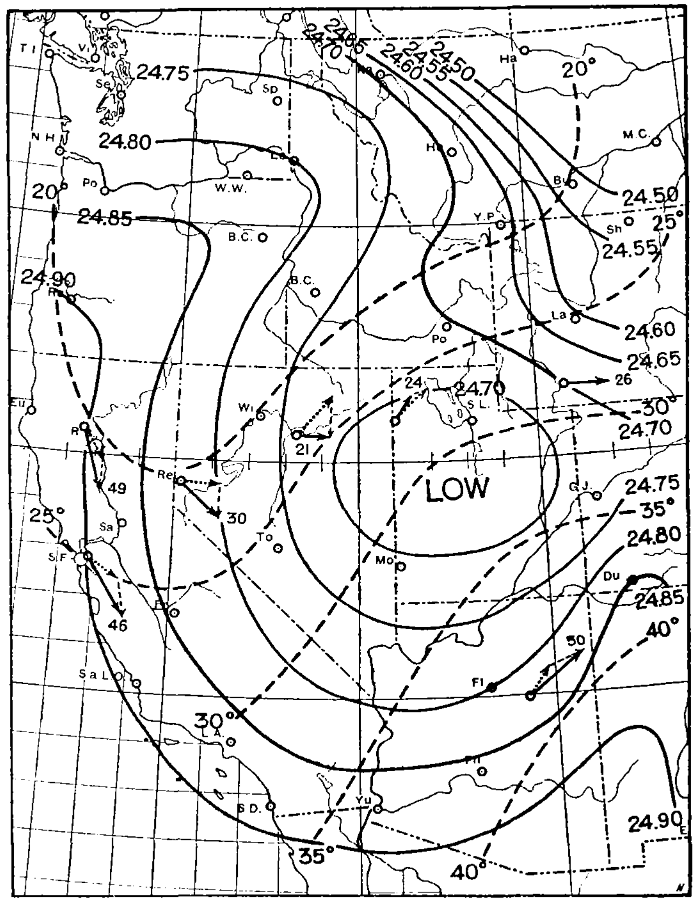

FIGURE 3.

When the pressure and temperature fields had been satisfactorily represented, the isobaric and thermal components for each of several points were scaled off. (The particular wind scales reproduced in figures 1 and 2 cannot be used on figure 3, because the scales are for a 1: 5,000,000 map projection while figure 3 is a different projection.) The vector representing the thermal component for each point was entered first, as an arrow with dotted shaft, always directed parallel to the isotherm passing through the point in question, with lower temperatures to the left of an observer standing with back to the wind. The vector representing the isobaric component, an arrow with broken shaft, was next entered for each point. It was in each instance placed with its tail beginning at the head of the thermal vector, and was directed parallel to the isobar passing through the point under consideration, with lower pressure on the left of an observer standing with back to the wind.

With the thermal and isobaric components thus represented, the resultant vector for each point was drawn from the tail of the thermal vector to the head of the isobaric vector. Drawn in this way the resultant vector gives the direction and speed of the wind at the point and altitude under consideration. In figure 3 the velocity represented by the resultant vector has been entered near the head of that vector.

In comparing the computed winds with the winds shown by the next scheduled pilot balloon observations it is interesting to note that the computed wind for Redding was north northwest 49 m. p. h. while the observed wind was north northwest 46 m. p. h. The computed winds for Winslow and Rock Springs were the same as the observed winds at those points at 11 a. m., E. S. T. and 5 a. m., E. S. T., respectively. The degree of accuracy indicated by the latter comparison is of course exceptional, and no such accuracy is claimed, nor could it be expected. If the time spread is considered, Rock Springs observation 5 a. m., Winslow observation 11 a. m., and computed winds for 7:30 a. m.—it becomes quite apparent that the true and computed winds were in all probability not identical. Nevertheless the figures on these three stations do indicate excellent agreement between the observed and the computed winds.

It is not possible to make a close check on the computed winds from Salt Lake City to California, due to the fact that no pilot balloon data were available in either the 5 a. m. or 11 a. m. collections. However, a rough check is possible from the trip log of a United Air Lines plane which flew from Salt Lake City to Oakland between 9 a. m. and 2 p. m., E. S. T. In the vicinity of Wendover, where the course lies almost due west, the log indicates a correction for a moderate quarter-headwind from the left. In that section of the route lying between Buffalo Valley and Reno the course lies roughly west-southwest. The log indicates a correction for a moderate quarter-headwind from the right near Buffalo Valley, and a moderate cross wind from the right near Reno. In view of the fact that the flight was made mostly on instruments, with light to moderate static reported, it does not seem probable that an exact course on the radio directional beam could have been maintained. Since the departure of only a small amount from the exact course would affect the wind computed from the ground speed and compass headings, a minute comparison between the winds so computed and those computed from the temperature and pressure fields would appear to be superfluous. However, the direction and approximate amount of the correction found necessary to navigate the course show satisfactory agreement with the winds computed from the pressure and temperature fields.

Before passing figure 3 it is interesting to observe that the wind at any level between the 5,000-foot plane and the 14,000-foot plane should be represented by a vector drawn from some point on the shaft of the thermal vector to the point of intersection of the isobaric and resultant vectors.

Similar methods can be used with the sea level pressure map in flat country, provided isotherms are superimposed on the pressure chart, representing the field of mean temperature applying between the surface and the altitude for which winds are to be computed. The scale reproduced in figure 2 has been prepared for use in conjunction with the sea level chart.

It will be interesting to some to know that the scales for the isobaric component may be used as geostrophic wind scales for the surface or sea level chart, and that if the corrections indicated for ΔP_0 and ΔT_m are applied, the scales so used will be highly accurate. In using the isobaric scale for this purpose, T_m should be taken as the surface temperature at the point where the wind is to be measured.

Theoretically, the wind scales should enable us to compute the winds aloft with a close degree of precision. In fact, from purely static considerations, the computations should be nearly exact. Difficulties will be encountered, however, and it is fitting that a few of them should be mentioned here.

Representation of the temperature field is likely to be the most common and most bothersome source of trouble. As indicated earlier, it is hoped that at centrally located airway meteorological offices there will be on hand sufficient temperature data to enable the meteorologist to obtain a creditable and satisfactory picture of the distribution of temperature at least once and possibly several times each day. Experience in drawing isotherms of T_m will gradually lead to greater accuracy than one can hope to attain at the first attempt. For example, it will be observed that in frontal zones the isotherms should be relatively crowded; that strong winds reported, although by only an isolated pilot balloon observation, usually indicate a relative crowding of the isotherms in that vicinity unless the amount of wind can be accounted for by the existing surface pressure gradient. This latter point holds so well that it is possible to determine the free air horizontal temperature gradient provided the surface pressure gradient and the free air winds are known.

If the temperature field is properly represented, the greatest source of error will have been eliminated. Smaller errors are always possible as a result of the fact that both the cyclostrophic and the isallobaric components of the wind flow have been omitted. We have assumed balanced, straight-line flow of the air. It is believed that the resulting errors will be very small and as a rule negligible, inasmuch as these components in the free air are small as compared to the speed of modern aircraft. Furthermore, the two components tend to act in the opposite directions and therefore to cancel one another: an isallobaric component to the left necessarily being associated with and bringing about a cyclostrophic component to the right, while an isallobaric component to the right should be associated with a cyclostrophic component to the left.

The influences of turbulence, friction, convection, and terrain (all closely related) have not been taken into account but should not prove too troublesome because the method here presented is designed to be used mainly for computing winds at altitudes where these are minor factors.

Finally, it should be emphasized that the method here suggested is not one for forecasting upper winds, but rather one for determining what upper winds should prevail at a given moment, past or present.

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