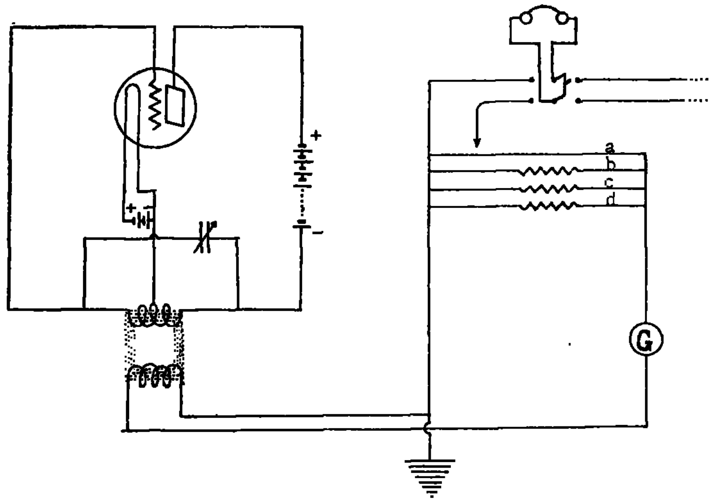

FIGURE 7.—Oscillation circuit for comparative audibility tests.

To illustrate, the values of $I \cdot \frac{ax}{ab}$ obtained for the averages of seven consecutive sets of readings were 0.059×10^{-7}; 0.065×10^{-7}; 0.0572×10^{-7}; 0.067×10^{-7}; 0.0556×10^{-7}; 0.055×10^{-7}; 0.053×10^{-7}. The greatest difference in the values obtained is between 0.067×10^{-7} and 0.053×10^{-7}, and assuming 0.067×10^{-7} to be the correct value we have an error of 20.8 per cent, in the reading 0.053×10^{-7}, as the amount of variation due to the ear. If the mean value of these readings is taken (0.0588×10^{-7}), we have 0.0098×10^{-7} as the greatest error, which gives a per cent of error equal to 14.6 per cent as the greatest variation. Since the human ear is extraordinarily sensitive at times and just as sluggish at other times, it would seem that the mean record should be the standard.

During the same period Arlington audibility was as follows: 4, 1, 22, 0, 2, 5.5; showing a fluctuation of audibility from 0 to 22 as compared with a fluctuation in the constant circuit from 0.0572×10 to 0.067×10^{-7} for the same dates. Or comparing the first two in each set of readings, constant circuit ratio 59:65, Arlington ratio 4:1, and any two consecutive readings on these two could be compared with like showings. Results show that while local noises or personal physical condition might have been the cause of the small variations in the audibility of the constant circuit, the same causes could scarcely have been the cause of the extreme variations in the carefully calibrated receiving circuit. Audibility of static for these seven days could be compared in the same way. It was 200, 500, 1,000, 200, 400, 200, 150.

ACKNOWLEDGMENTS.

Credit is due Prof. J. C. Jensen, who, in 1916 or earlier, first planned this work and has since directed its progress. Credit is also due him for taking the series of readings on 9YT, Wayne Normal (fig. 5), and for taking a series of readings with the observer by which a ratio was obtained between their audibilities on the same station under similar conditions, this ratio being used when it occasionally becomes necessary for him to take readings for the regular observer. The comparison method for measuring sound intensities, as shown in Fig. 7, originated with Dr. J. M. Miller of the Bureau of Standards and was suggested to us by Mr. L. E. Whittemore of that Bureau. Through the courtesy of C. F. Marvin, Chief of the Weather Bureau, the large daily weather maps were made available for our use.

THE CONNECTION BETWEEN PRESSURE AND TEMPERATURE IN THE UPPER LAYERS OF THE ATMOSPHERE.

By W. H. DINES, F. R. S.

[Benson, Wallingford, England, Aug. 30, 1922.]

The large number of observations that have been made over Europe on the temperature of the air at heights reaching up to 15 or 20 kilometers have shown that at certain heights the two most important variables of the atmosphere, the pressure and the temperature, are most closely connected with each other. In the following remarks the nature of this connection is set out as far as possible in nontechnical language, and a suggestion is made as to the way in which the connection is brought about by the ordinary physical laws.

The careful statistical examination of the records obtained from the ascents of many hundred registering balloons in the British Isles and on the Continent has disclosed the facts given below. Let us use P_n to denote the pressure at a height of n kilometers above sea level and T_n for the corresponding temperature, also $T_{n,m}$ for the mean temperature with regard to height between n and m kilometers. Let δP denote the departure of P from its mean for the geographical position and the height and δT the corresponding quantity for the temperature. Then at the surface there is found to be no connection between δP_0 and δT_0, if the barometer is high so that δP_0 is positive there is no tendency for the temperature to be either above or below the normal value. At 1 kilometer, however, a bias is beginning to appear, if δP_1 is positive there is a distinct tendency for δT_1 to be positive also and if δP_1 is negative δT_1 is probably negative also. This tendency strengthens with increasing height until over the layer from 4 to 8 kilometers δP and δT will almost certainly have the same sign and will be almost strictly proportional to each other. Above 8 kilometers the connection between the quantities falls off rapidly until at about 10.5 we get back to the conditions that prevail at the surface. There is no connection. But above 10 kilometers the conditions change and the δT_{10} changes sign, so that at 11 kilometers and upward the positive δP_{10} are associated with the negative δT_{10}, and where the pressure is above its normal value the temperature is below, and vice versa.

TABLE 1.—*Correlation coefficients.*

	T_0 and P_0.	T_1 and P_1.	T_2 and P_2.	T_3 and P_3.	T_4 and P_4.	T_5 and P_5.	T_6 and P_6.	T_7 and P_7.	T_8 and P_8.	T_9 and P_9.	T_{10} and P_{10}.	T_{11} and P_{11}.	T_{12} and P_{12}.	T_{13} and P_{13}.
January to March	−0.02	0.54	0.82	0.79	0.86	0.85	0.84	0.87	0.91	0.81	0.35	−0.32	−0.38	−0.37
April to May	.14	.28	.49	.79	.89	.89	.92	.87	.81	.45	.20	−.12	−.24	−.01
June to September	−.02	.31	.56	.72	.75	.81	.83	.87	.87	.88	.43	−.08	−.41	−.19
October to December	.33	.56	.76	.77	.83	.87	.85	.85	.86	.78	.29	−.24	−.34	−.50
Means	.11	.42	.66	.77	.84	.85	.86	.87	.86	.71	.32	−.19	−.36	−.28

These figures are taken from *Geophysical Memoirs* No. 13 (M. O. 220c), published by the Meteorological Office, where many other correlation coefficients are given in detail.

This general relationship between the pressure and the temperature was discovered many years ago by Teisserence d'Bort. It is simply the statement that in the cyclone the troposphere is cold and the stratosphere warm and in the anticyclone the troposphere is warm and the stratosphere cold; the remarkable point about it is the extreme closeness of the connection from 4 to 8 kilometers height. To those accustomed to deal with statistical data by means of correlation this is plainly brought out by the table of correlation coefficients given above. The values are obtained from some 200 observations made in the British Isles. Very similar values hold for the Continent and for Canada. The coefficients are uncorrected for the observational errors and are on that account about 5 per cent too low. Thus from 5 to 8 kilometers the correlation reaches the very high value of 0.90. Since the square of the correlation coefficient is the measure of the influence exerted by one of the quantities upon the other, in this case if we take the variation of temperature as being due to the pressure changes it appears that four-fifths of the variation is so produced leaving only one-fifth for all other causes combined, such as the direction of the wind, the presence or absence of cirrus cloud, etc. It may be added that except in the first 1 or 2 kilometers there is no correlation between the temperature and the direction of the wind.[1]

There is another connection between temperature and pressure that must be noticed. The fall of temperature with height, the "lapse rate," as it is called in England, ceases in latitude 50° N. at about 10.5 to 11 kilometers and the height at which it ceases is commonly denoted by *Hf*. This height varies for England from 7.5 in a deep cyclone to perhaps 13.0 kilometers in an anticyclone. It varies with the surface pressure but it varies much more closely with the pressure at 9 kilometers, the correlation deduced from many hundred observations in England and on the Continent being as high as 0.84. The pressure at any height *hs* is calculated by Laplace's formula from the pressure at the surface and the harmonic mean temperature of the intervening air column, and it has been contended by some that the high correlation is due to this fact. It is necessary, therefore, before seeking further explanation to examine this contention. Keeping the height constant and differentiating Laplace's formula we obtain an equation of the form

$$\delta P_n = A\delta P_o + b\delta T_{n,o} \qquad (1)$$

In this equation a and b are constants which depend upon the height, n kilometers, and the mean values of P and T. Obviously for determining the pressure at a small height it is the term involving a, the surface pressure, that is important, whereas for great heights it is b that matters. Using millibars and degrees C. for units a and b become numerically equal for a height lying between 2 and 3 kilometers, while for a height of 10 kilometers, b becomes five times as great as a. The inevitable relationship shown in equation (1) ensures therefore that there will be a very high correlation between P_n and $T_{n,o}$ unless the height n is small and that the correlation will increase with increasing height; it is apparent also that the coefficient b is essentially positive and that equation (1) must produce a positive correlation. The term T_n is not directly involved in (1) but it is so indirectly because it is used in forming the mean value $T_{n,o}$, but unless the air column is short P_o alone is relatively unimportant. Hence the relation (1) does produce some small correlation; that it is utterly incapable of causing the close connection shown by the observations the following consideration shows.

[1] See *Geophysical Memoirs* No. 2, M. O. 210b, p. 45, and *R. Met. S. J.*, Vol. XLVII, No. 197, Jan., 1921, p. 26.

It has been proved above that the relation $\delta P_n = a\delta P_o + b\delta T_{n,o}$ produces (1) a positive correlation (2) a correlation increasing with the height. The observations show a negative correlation in certain parts which is contrary to (1) and a decreasing correlation from 8 to 13 kilometers which is contrary to (2). The connection between P and T can not therefore be due to Laplace's formula.

The second relationship between pressure and temperature is given by the equation

$$\frac{\delta T}{T} = 0.29\frac{\delta P}{P} \qquad (2)$$

This is the well-known connection between small variations of pressure and temperature that occur when air expands or contracts under circumstances such that it can neither gain nor lose heat. The coefficient 0.29 refers to dry air; for expanding air in which the cooling is producing the condensation of water vapor the value is lower. The requisite condition that there shall be no gain or loss of heat is met with in air not in contact with the surface of the earth that is altering its pressure quickly, say, in a time conveniently measured in hours. If the change takes days, it cannot be called adiabatic, owing to the effect of radiation. Let us assume for the present that no condensation is occurring and that the changes are adiabatic.

The assumption that air follows the rule of changing 10° C. per kilometer change of height is not so far from the truth as might at first sight appear. For the rule applies practically to all descending air since the amount of water that air can contain without spilling it as rain is very small. Secondly the space where vapor is condensing compared with the atmosphere up to, say, 20 kilometers is also small. If we take rain as the index, rain occurs on the average perhaps 1 hour out of 20 and the rain cloud may be about 2 kilometers thick. This gives $\frac{1}{20}$ of $\frac{2}{20}$ or one-half per cent as the space in which the rule of the dry adiabatic rate does not hold. If we take cloud as the index, we may estimate as follows. It has been held that half the sky is on the average covered with clouds and we may take the average thickness as 1 kilometer. This is $\frac{1}{2}$ of $\frac{1}{20}$ or $2\frac{1}{2}$ per cent. Thus even supposing we consider the part of the atmosphere below the 10° kilometer level above the percentage is still only 1 (in 20) or 5 per cent. Strictly the rate for dry air does not quite reach 10° per kilometer, it is 9°.84, but for ordinary air so long as condensation does not occur the value of 10° per kilometer is a sufficiently close approximation.

In equation (2) T must be measured from the absolute zero, but the pressure may be in any units. Taking P in millibars, the value of 0.29 T/P for any height is obtained by substituting the mean value of T and P for that height. The values for England at any exact kilometer height up to 20 so obtained are given below, δP being taken as 1 millibar.

TABLE 2.

Height (km.)	0	1	2	3	4	5	6	7	8	9	10
C	0.08	0.09	0.10	0.11	0.12	0.14	0.15	0.17	0.19	0.22	0.25

Height (km.)	11	12	13	14	15	16	17	18	19	20
C	0.29	0.33	0.39	0.45	0.54	0.62	0.73	0.85	0.99	1.15

The figures in this table express in another form the fact that dry air rising adiabatically for 1 kilometer will *fall 10° in temperature, for if* 10 be divided by the decimal giving δT for any height the quotient will express the approximate change of pressure per kilometer in millibars at that height. But the change of pressure may be produced by any means. It is not necessary that the air should rise or fall. Change of position without change of level, if the air comes under a new pressure is, equally efficacious.

The connection between pressure and temperature changes shown algebraically by (2) and set out numerically for each height in Table 2 tends toward a positive correlation increasing with height between P and T. The actual connection as disclosed by some 200 observations is set out below in Table 3.

TABLE 3.

$\delta T_0 = 0.04\delta P_0$	$\delta T_4 = 0.42\delta P_4$	$\delta T_8 = 0.44\delta P_8$	$\delta T_{11} = -0.11\delta P_{11}$
$\delta T_1 = 0.19\delta P_1$	$\delta T_5 = 0.46\delta P_5$	$\delta T_9 = 0.31\delta P_9$	$\delta T_{12} = -0.29\delta P_{12}$
$\delta T_2 = 0.34\delta P_2$	$\delta T_6 = 0.48\delta P_6$	$\delta T_{10} = 0.14\delta P_{10}$	$\delta T_{13} = -0.28\delta P_{13}$
$\delta T_3 = 0.39\delta P_3$	$\delta T_7 = 0.47\delta P_7$		

These figures are obtained by the ordinary statistical method. *They are, in fact, regression equations between* P and T at each height, thus $\delta T_6 = .48\delta P_6$ means that taking the average of many observations where the pressure at 6 kilometers height is 1 millibar above its average value the temperature will be half a degree (0°.48 C.) above its average. The height is only carried to 13 kilometers because the observations above that height are not sufficiently numerous to give reliable values.

On comparing Tables 2 and 3 it will be seen at once that they do not agree. The observational results shown in Table 3 prove that above 10 kilometers a high pressure is usually associated with a low temperature, over an anticyclonic region the stratosphere is cold, whereas dynamic warming as shown in Table 2 should produce a high temperature, for a high pressure at the ground level remains a high pressure up to about 20 kilometers.

But Table 2 is founded on the assumption that the air found at any level had originally, i. e., before the change of pressure, the temperature and pressure corresponding to that level and this assumption is not warranted. Take the particular values at 6 kilometers as an example. The mean pressure there is 469 mb. and the mean temperature is 248°. If these conditions held and the pressure then rose to 489 mb. the temperature in consequence would rise to 251°. But if instead of a change of pressure occurring at 6 kilometers the pressure both at 6 and 7 kilometers had remained the same, but air for some reason had been forced down from the 7 to the 6 kilometer level, the rise of temperature would be 10°; and since the mean at 7 kilometers is 241°, on reaching the level of 6 kilometers its temperature would be 251°, as before. Similarly for other levels. Thus, without impugning the truth of equation (2) which, since it is a well-established law of physics, we may not do, we can explain the observational results by assuming a vertical motion in the atmosphere dependent on the height and on the distribution of pressure. It is strange that the effect upon the temperature of a small vertical component in the motion of the air should have received so little attention from meteorologists. Hann more than 20 years ago explained the warmth of high Alpine peaks during anticyclonic weather in this way but his suggestions have not been followed up. It is the more curious because the fact that rain is due to the dynamic cooling of an ascending current is emphatically stated in most modern textbooks on meteorology.

It is stated that rain is due to the "cooling" of the air, and yet the cooling of an ascending current is not considered except in the special case when rain is produced. All air that ascends in one place must come down again somewhere else and doubtless there are plenty of air currents with a rapid horizontal component, but with a small vertical component also so that they are blowing at a small angle to the horizontal plane. If this angle be only three minutes of arc its effect on the temperature is equal to that due in a south or north wind to the change of latitude; and yet this small inclination even in a thick, rapidly-moving current would hardly produce a measurable quantity of rain in an hour.

The above statement refers to England where the change in temperature with latitude is small and is the result of a simple calculation.

The change of temperature from Equator to pole is close to 45° C.; that is, 1° C. for 2° of latitude, or for a distance of 222 kilometers. The lapse rate of temperature is 6° per kilometer height and the adiabatic lapse *rate is 10° per kilometer for dry air, hence air rising* 250 meters will find itself 1° C. colder than its new surroundings. Thus a south wind with an upward gradient of 0.25 in 222, or say 1 in 900, which represents an inclination of less than 4′, will on the average, if no vapor is condensing, find that the two causes for a temperature different to its surroundings just cancel each other and it will show no sign of its place of origin. If then we have vertical motion in the atmosphere, we have a cause of change of temperature at least as powerful as an equatorial or polar wind and it seems certain that currents with an inclination to the horizon much over 4′ are quite common. The rate at which cyclonic rain falls proves this because at the ordinary temperature of the rain-producing strata for central Europe, say 0° C., at 2 kilometers height, the water can not be provided unless there is a fairly rapid upflow of saturated air.

Let us estimate the slope necessary to produce the quite common rate of rain of 2.5 mm. per hour. Take a current of saturated air at 5° C. blowing up a slope of 1 in 100. Consider a cube of 1 cubic meter volume passing up 1 kilometer of such a slope. Using Hann's figures it will fall 0.06° C. in temperature and in consequence condense about 0.025 grams of water. The water will be spread over 1,000 square meters of ground and will provide 2.5×10^{-8} mm. of rain. We have therefore to allow for 10^8 cubic meters per hour passing over each square meter of ground. By assuming the current to be 2 kilometers thick the factor is reduced to $5 \times 10''$ and the velocity of the current required is 50 kilometers per hour. We can hardly take the velocity as much over 50 kilometers per hour or its thickness as more than 2 kilometers; even if we took a thickness of 4 kilometers it would not add much to the rainfall because the upper part would be too cold to carry much moisture. Hence the slope of rain-bringing winds can not be much less than 1 in 100, and the cooling or heating effect of the rise or fall of air in such a wind is some 10 times as great as that due to change of latitude in a north or south wind. Since ascending currents in one place must be compensated by descending currents, bringing down an equal mass of air elsewhere, we have direct proof of vertical motion in the atmosphere sufficient to explain all the temperature anomalies that occur.

The dependence of the temperature upon the pressure can be readily explained if we admit the vertical motion of the air; there is no rigid proof that the explanation is the right one, but it fits in with all the known facts of the case.

If, in the laboratory, we arrange an endless tube with two vertical parts and warm one part, continuous circulation of the air in the tube would set in and the air in the vertical part that is being warmed would be rather warmer than the air in the other parts. But suppose we could perform the experiment in the open air with the vertical branches some few kilometers instead of some few meters in height, the result would be quite different. Unless the heat supplied were very considerable, continuous circulation would not set in. About half the heat supplied to one branch would appear immediately as sensible heat in the other branch and the mean temperatures in the two vertical parts would both rise equally. This result is due to what Sir Napier Shaw calls the resilience of the air. This resilience resists any vertical motion, and is due to the fact that the ordinary lapse rate of temperature is some 4° C. per kilometer less than the dry adiabatic rate.

Let us take, then, an imaginary tube $A\ B\ C\ D$ with its walls impervious to heat, suppose A and C to be vertical parts of a few kilometers length and B and D to be horizontal branches of 50 or 100 kilometers, and suppose the air inside to be at the mean atmospheric temperature for the corresponding height. Suppose now that the air is compelled by some external force to shift by 250 meters round the tube in the direction $A\ B\ C\ D$ so that the air at A originally will come to a, a point 250 meters above A, and the air at C originally to c, a point 250 millimeters below C. The result on the temperatures will be that save near the corners the temperature in the column A will be 2° C. below that in the column C for points at the same height because the lapse rate being 6° per kilometer and the dry adiabatic rate 10° per kilometer, the air at any point A on the rising side will be 1° cooler than before and at any point C on the falling side it will be 1° hotter. Now suppose the force causing the displacement to cease; the air in the column C being the warmer will rise and the initial condition of the same temperature at the same height on both sides will be restored. It is not necessary to postulate any force at all. Starting from the initial conditions, if by any means the air in one vertical branch becomes hot or cold, the resulting circulation which sets in under gravity continues until equality of mean temperature in the two vertical branches is attained.

The force that has been supposed to act may well be due to the distribution of pressure and it has been shown that if a suitable external force is acting on the air differences of temperature at the same level may be maintained but that if no force save gravity is acting, heat applied at a point A will not have the sole effect of warming the air at A but will, owing to the circulation that will ensue, also warm the air at the same level within a moderate distance of A.

If A be near the ground, vertical circulation can not readily occur and we find, as a matter of fact, that near the ground there is no appreciable correlation between pressure and temperature, but in the upper half of the troposphere, where vertical motion is least impeded by proximity to the boundaries, the correlation is closest.

Now suppose a tube of flow like $A\ B\ C\ D$ to be moving with the general air current and to be so placed that its horizontal branches more or less coincide in direction with the line of the isobars. There is not likely to be any appreciable force due to the pressure distribution to cause circulation. Hence, as already shown, there should be equality of temperature for points at the same height. That is to say, since the points are approximately over the same isobar, at points at the same pressure, and this is the first requirement to produce a high correlation between pressure and temperature.

Next suppose our imaginary tube of flow to lie with its horizontal branches more or less at right angles to the isobars or to the current of air in which it is moving. There is now a pressure gradient along both horizontal branches and it seems quite likely that integrating round the tube there may be an effective force causing a shift of air. Suppose it to cause such a shift in the direction $A\ B\ C\ D$; then, as has been shown, the air in the vertical branch A will be colder than that in the vertical branch C and will continue colder while the force causing the shift is in action. Suppose A to lie inside so that it is nearer the cyclonic center than C. Then we have lower temperature combined with lower pressure and hence a positive correlation.

If we could experiment with an actual tube, it would not matter in what part of the tube the force was applied, but the tube is only a convenient mathematical conception. If, however, the circulation will occur in the tube, much more will it occur in the free atmosphere, and it is obvious from what has been said that if a force acts horizontally in the atmosphere so as to shift air from a region A to a region B, A and B not being too near the earth's surface, the result is a tendency to cool the air under A and over B and to warm the air over A and under B. This follows because the deficiency of air at A caused by the flow must be made good and the excess at B disposed of. Remembering the great disproportion between the vertical and the horizontal scale that any representation of a cyclone on paper must present, we see that the deficiency as well as the excess would naturally be compensated by a vertical component of the air's motion.

Now the distribution of temperature, we have to explain, is a high temperature in the stratosphere over a cyclone and a low temperature in the troposphere, most marked in the upper part; that is to say, a high temperature above 10 kilometers, a low temperature occurring with very great certainty from 4 to 8 kilometers and a low temperature, but not so certainly, from 1 to 4 kilometers. These conditions, which many hundred observations in Europe have shown to exist beyond dispute, are fully explained dynamically if we suppose the upper winds to produce a sort of sucking action on the air at about 9 kilometers height, so as to take the air from the upper part of the troposphere; for this motion, as already shown, produces the required temperatures. This also explains the fall in height of the boundary between the troposphere and stratosphere which occurs over every well-marked cyclonic area, for as the column of air falls its lapse rate remains comparatively unaltered, all the temperatures being raised by about the same amount, so that the usual inversion at the boundary still remains an inversion, but at a higher temperature and lower level.

The statement made above with regard to a cyclonic area refers with equal truth to an anticyclonic area if the terms "high" and "low" be interchanged. The horizontal flow of air at 8 or 10 kilometers height toward an anticyclone will press up the upper boundary of the troposphere and cause cold above in the stratosphere. It will also press down the air under the locality where it ends and cause warmth in the troposphere.

The only difficulty I see in this explanation is the difficulty of assigning a reason for the flow of air at 8 to 10 kilometers from the cyclonic to the anticyclonic area. This is the height of the strongest winds, and it is easy to ascribe the flow in general terms to the geostrophic and centrifugal action of the wind, but that does not carry us much farther. If it is correct, it means that cyclones are generated in the upper part of the troposphere by the general circulation of the atmosphere and spread downward. Also that there is a slight outflow of air from the cyclone above; it need only be slight, just as there is a slight inflow below near the ground.

A correct theory of the formation and propagation of cyclones must be able to account for the well-marked distribution of temperature in the upper air that accompanies them and the temperature is almost certainly due to dynamic heating and cooling. The changes of temperature accompany the changes of pressure and are large from day to day, far larger than would be possible if they were due solely or chiefly to radiation. It is natural to look to the direction of the wind to account for changes of temperature, but the statistical evidence is perfectly distinct in showing that the direction of the wind, save quite close to the earth, has only a trifling effect upon the temperature. Above 2 kilometers there is no appreciable correlation between the south to north or west to east components of the wind and the temperature. This is the case if the actual surface wind, the *gradient* wind, or the drift of the balloon which carries the instruments be used. Doubtless, as Capt. C. K. M. Douglass urges (R. Met. S. J., Vol. XLVII, No. 197, p. 23), it is the place of origin of the air, not its temporary direction that matters, but in view of the absence of correlation between wind direction in the upper air and temperature it does not seem possible to me that cyclones should be caused by the action of polar and equatorial currents. On the other hand there can be no doubt that the mild winters of western Europe are due to the prevalent southwest and west winds coming from the warm waters of the north Atlantic.

In the suggested explanation of the correlation between pressure and temperature nothing has been said about the time requisite for the adjustment. If the explanation is to be feasible, the time required must be comparatively short; otherwise radiation would prevent the changes of temperature from being adiabatic. If we neglect frictional resistances, it can be shown by elementary dynamical considerations that equalization of temperature between places 200 kilometers apart would take about an hour. The time required varies as the square root of the distance, so that the equalization between two places on the ordinary weather chart is only a matter of an hour or two, in which time radiation would not have much effect. But the assumption that frictional resistances may be neglected is certainly a large one owing to the eddy viscosity of the atmosphere. However, the gradient wind appears to adjust itself with considerable rapidity to the distribution of pressure notwithstanding the eddy viscosity, so perhaps the time is not greatly increased by the same cause. But the retardation due to eddy viscosity will vary as the distance, so that for large distances it may be very considerable. This may explain why differences of mean temperature exist in winter between, let us say, England and eastern Europe at a few kilometers height, though it may be noted that such differences are far smaller than those found at ground level.

AVERAGE FREE-AIR WINDS AT LANSING, MICHIGAN.

C. L. RAY, Observer.

[Weather Bureau, Lansing, Mich., Oct. 14, 1922.]

The first pilot-balloon ascension at this station was made June 10, 1919, and flights have been made daily since that time, except when impossible through inclemency of the weather. Flights were made at 7 a. m. and 3 p. m. until August, 1921. Beginning with the flight of August 1, the morning ascensions were discontinued, and the single flight daily at 3 p. m. has been made since that time.

The results set forth in this paper have been based on the flights made during the three-year period, June 1919–May 1922, inclusive. The number of flights obtained during that time, and listed by seasons and altitudes, follows:

TABLE 1.—*Number of pilot-balloon ascensions, June 1919, to May, 1922, inclusive.*

Altitude.	Spring.	Summer.	Autumn.	Winter.	Annual.
m.					
Surface	364	420	366	324	1,474
250	364	420	366	324	1,474
500	344	414	353	286	1,397
750	323	401	326	242	1,292
1,000	307	394	290	213	1,204
1,500	264	359	254	170	1,047
2,000	231	313	226	143	913
2,500	194	266	177	127	764
3,000	164	235	148	110	657
3,500	136	190	116	87	529
4,000	117	170	94	76	457
4,500	103	152	66	56	377
5,000	85	136	56	40	277
6,000	52	100	34	27	213

The percentage of winds from various directions over this three-year period is shown in Table 2, and as will be noted shows over 50 per cent of the surface winds with a south component and more than 56 per cent with a west component. At 4,000 and 6,000 meters elevation the preponderant direction lies between west and northwest, and slightly favoring northwest. The detailed percentages are as follows:

TABLE 2.—*Percentage frequency of winds observed from various directions.*

Meters.	N.	NNE.	NE.	ENE.	E.	ESE.	SE.	SSE.	S.	SSW.	SW.	WSW.	W.	WNW.	NW.	NNW.
Surface	6	4	5	3	3	3	5	5	9	10	11	8	10	6	8	4
1,000	5	5	4	2	3	2	2	3	5	6	10	11	16	10	11	5
2,000	6	3	4	2	2	2	2	1	2	4	8	12	18	14	12	8
4,000	4	3	4	2	1	1	1	2	2	4	8	7	17	19	16	9
6,000	6	1	4	1	0	3	1	2	1	4	6	8	15	20	21	7

The information contained in this table has been used in the graphical representation (fig. 1), which shows probably to better advantage the results obtained. Above 2,000 meters west to northwest winds generally prevail.

In Table 3 are given the mean free-air winds for the different seasons and the mean annual directions and velocities. Southwest winds prevail at the surface during the spring, summer, and *autumn months*, giving place to a west direction in the winter season. All surface velocities are close to three meters per second