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REPORT ON AN EXPERIMENT IN FIVE-DAY WEATHER FORECASTING

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INTRODUCTION

The following report is presented as a statement of progress made at the Massachusetts Institute of Technology (M.I.T.) in the investigation into the possibility of extending the range of reliable weather forecasts. This project has been supported at M.I.T. and other private institutions by Bankhead-Jones appropriations since September, 1937. This report is concerned only with the work completed or in progress at M.I.T. The complementary program now in progress at the Weather Bureau in Washington is referred to only in so far as it has contributed directly to these investigations.

Furthermore, the following report refers only to the last two years of the M.I.T. project. The first year of the three-year project was given over principally to the study of the results obtained by long range forecast methods already in use, and to the establishment of a northern hemisphere synoptic weather map procedure as a necessary precedent to the preparation of weekly forecasts on a synoptic basis. The results of the M.I.T. study of certain long range forecast methods already in practice are included in a general survey of such methods already published.* The synoptic charts prepared at M.I.T. during that first year of the investigation are listed in an appendix to this report, together with those of the last two years. The preparation of weekly forecasts carried on during a part of that first year was so experimental in nature, and the procedure was so much changed the following year, that the results obtained were considered neither sufficiently significant nor comparable enough with the later forecast results to merit any discussion.

The present report is divided into three principal sections.

Section I presents in condensed form our present conception of the essential nature of the general circulation, and discusses briefly the background of one or two of Professor Rossby's theoretical considerations concerning the general circulation which have found statistical and synoptic application in this investigation.

Section II contains in brief form the results of synoptic and statistical checks of a large number of hypothetical relationships which might be assumed to hold in the earth's atmosphere. These include possible relationships in the large scale features of the general circulation, relationships between the general circulation and its different branches or centers of action, between the different branches or centers of action of the general circulation, between characteristics of the general circulation or its branches and anomalies of the meteorological elements in certain regions, between anomalies of the meteorological elements in one region and those in another region, and even between solar activity (sunspots) and characteristics of the general circulation or anomalies of the meteorological elements. The aim was to investigate possible interrelationships of all kinds, either with or without lag, in order to detect as many interaction principles or points as possible in the earth's atmosphere, whether they had direct or only the most indirect bearing on the forecast problem. The relationships investigated applied to daily, weekly, monthly, seasonal, or annual mean conditions. They were selected for investigation either from theoretical or practical considerations of the nature of the general circulation as outlined in Section I, or on the basis of popular beliefs which have long been current among

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* *Monthly Weather Review*, Supplement No. 39, 1940.

meteorologists, or on the basis of direct observation of data which looked promising. The majority of these hypothetical relationships are found to be quite weak when subjected to rigid statistical checks, but all such results, whether positive or negative, are summarized in this report.

Section III outlines the five-day forecast routine practice which has been carried on at M.I.T. during the greater part of the past two years on a weekly basis. It includes a statistical analysis of the verification results.

In the conclusion are summarized the results of the investigation which thus far appear significant enough to justify their consideration in five-day or longer range forecasts. Suggestions are offered as to further steps which might profitably be taken if the investigation is to be continued.

Finally there is an appendix in which are listed all the daily synoptic maps and mean charts and diagrams of surface and upper air data which have been plotted and analyzed at M.I.T. in connection with this project during the past three years. The importance of such a list is apparent when it is realized that inevitably in an investigation of this kind much the greater part of the time and effort expended is consumed in the routine or semi-routine duties involved in the preparation of such charts.

SECTION I. GENERAL CIRCULATION OF THE ATMOSPHERE

A. INTRODUCTORY REMARKS

The physical laws which govern the motion of the atmosphere and the changes which occur in the moving air masses are known in principle. These laws are expressed in condensed form by the fundamental differential equations of hydrodynamics and thermodynamics.

The problem of deriving the laws of atmospheric motion in explicit form is, however, exceedingly difficult. The principal reason is that the atmosphere is much more complex than the "fluids" generally discussed in hydrodynamics. The atmosphere is not only heterogeneous and of variable composition due to condensation and evaporation, but it is also a "thermally active" fluid continually receiving or losing heat. This atmospheric heat exchange is maintained by the processes of radiation and condensation, both of which are determined by the moisture distribution (see p. 10). The moisture distribution depends upon and is modified by the motion of the atmosphere. The motion, on the other hand, is primarily caused by the thermal action of heating and cooling, which is determined by the moisture distribution. This double interdependence between the motion and the moisture distribution leads to unsurmountable mathematical difficulties in theoretical analysis of atmospheric motion. For the gaining of some insight into the dynamics of the atmosphere two different approximative methods have therefore been adopted in dynamic meteorology.

In strict theoretical analysis which constitutes the field of dynamical meteorology proper, it is generally assumed that the heat supply is a known quantity and that the humidity is individually constant. These assumptions lead to a system of differential equations which give a rough approximation to true atmospheric conditions. However, even these simplified equations are so complicated that no general theory has yet been developed for their solution. The procedure has therefore been to select relatively simple "models," i.e., systems of fluid motion which on the one hand may be investigated by the differential equations, and on the other hand give a summary representation of some real atmospheric motion. The "models" are from the beginning selected from steady state motions of the circulative type, since such systems are not too difficult to discuss, and since they resemble certain types of atmospheric motion. Other models are in the nature of small oscillations or waves, superimposed on the steady state circulations. A particularly important motion of the oscillatory type is the young wave cyclone which is observed in the polar front on synoptic weather maps. Systematic study of small oscillations has shown that instability frequently occurs in the atmosphere, especially in the form of small-scale "eddies" of varying form and magnitude. An important question is then whether the mean motion which remains when the eddies are eliminated can be satisfied by equations similar or identical with those valid for "laminar flow." Only then will it be permissible to apply theoretical considerations for the discussion of the "turbulent" atmospheric motion.

Through intensive studies during the last twenty-five years a systematic theory of turbulent motion has started to develop. This field, which has been given the somewhat misleading name "Fluid Mechanics," was initially dealing with problems in aeronautical and hydraulic engineering. Only recently, mainly through the fundamental work by

Rossby, have the principles of fluid mechanics been adapted to atmospheric hydrodynamics. But the theory is here in the early stage of development, and it is still an open question whether the results derived from the purely theoretical models can be applied to real motion in the atmosphere.

The second method of approach aims at a direct study of the real motion of the atmosphere. In order to render this study possible, it is necessary to construct models with nature as the starting point. These models which summarize what we really know from observations are too complicated for a strict theoretical analysis. But it is generally possible to simplify them sufficiently to make them accessible to elementary reasoning based on the fundamental differential equations. The problem is to "understand" the dynamical properties of the models, i.e., to follow their evolution by means of the principles laid down in the general equations.

This branch of dynamical meteorology which deals with the study of more "real" models by mixed theoretical-empirical methods, provides the basis for the theory of "the general circulation of the atmosphere." The immediate purpose of such a theory is, on the one hand, to establish what the actual motion of the atmosphere is, and on the other hand, to understand the action of the determining dynamical factors underlying the variation from one weather situation to another. Only with the aid of a rational theory of the general circulation can the problem of long range weather forecasting be attacked with any hope of real success.

No complete theory is by a long way available as yet. This is due on the one hand to the fact that the empirical knowledge of the state of motion of the atmosphere and its changes is incomplete, so that the models by which the atmospheric conditions are summarized only are rough approximations, and in some cases even may be in conflict with real conditions. On the other hand, our dynamical "understanding" of the models is in many respects only qualitative, and the relative importance of the various dynamic factors is often difficult to estimate.

The research project which has been undertaken at M.I.T. during the past four years has had as its fundamental guiding principle the search for more accurate empirical models of the real atmospheric conditions. In this work synoptic analysis and statistical study of a world-wide observation material have been undertaken. The empirical models thus derived have been adjusted to accord with general physical and dynamical principles, through frequent theoretical "checks" of the models. And new theoretical results derived from these checks have in turn guided the further empirical search for more reliable models. Thus there has been a continual mutual cooperation between the synoptic-statistical investigations on the one hand and theoretical research on the other.

The aim of this general introduction to the synoptic-statistical report is to give a brief and general outline of the present state of our dynamical understanding of the atmosphere, and to point out what has been added to this understanding as a result of this research project.

B. The Conservation Principles in the Atmosphere

The most important of all real atmospheric models is that of the mean circulation of the whole atmosphere when all diurnal, seasonal and annual variations are eliminated. When this model is approximately known and an understanding of its dynamical properties has been obtained, the next problem will be to determine the *regular* annual and seasonal variations of this circulation and to attempt to estimate the relative significance of the dynamical and thermal factors which are responsible for these variations. Only when this has been accomplished will it be possible to approach the more difficult problem of the irregular variations from one weather situation into the next. The dynamical understanding of these irregular variations is an essential requirement for a rational method of "long-range weather forecasting."

In this discussion only the mean circulation of the atmosphere will be considered in some detail. The model of this mean circulation is idealized in so far as to neglect zonal variation in the meteorological elements arising from the asymmetric distribution of continents and oceans. The result will then be a certain motion symmetric with respect to the axis of the earth, and symmetric with respect to the equatorial plane. Some concluding remarks will then be made concerning the modification of this symmetric model arising from the asymmetry of the land and water distribution.

The intensity of this idealized circulation is determined empirically from the mean values of the meteorological elements wherever such data are available. Where some of the elements are missing the model is adjusted to accord with the general relation which must exist between the various meteorological elements according to the equations of motion, or according to principles derived from these equations with the aid of some simple theoretical model. Still over large regions of the atmosphere, particularly at high levels, all direct observational information is completely lacking. To overcome this difficulty we make use of the so-called conservation principles of which the following three are most important:

(1) The total amount of heat contained in the atmosphere cannot change from year to year; therefore on the average the atmosphere must give off the same amount of heat as it receives.

(2) The total mean amount of kinetic energy of the atmosphere is constant. This simply means that mean circulation as defined above is constant, and hence all the heat which through the circulation is converted into work, and hence into kinetic energy, must be dissipated again through friction.

(3) The mean circulation must adjust itself in such a fashion that the stresses exerted on the surface of the earth by the wind systems of the circulation have a total moment zero with respect to the axis of the earth. If this were not the case the earth would speed up the rotation of the atmosphere, and the atmosphere would slow down the rotation of the earth, or vice versa. Hence the moments of the east winds at the surface must be balanced by the moments of the west winds. These three principles are helpful in the final construction of the mean circulation model.

C. THE HEAT BALANCE OF THE ATMOSPHERE*

First the earth and the atmosphere will be considered as one system. The exclusive external source of energy of this system is the solar radiation. This radiation has been determined through extensive measurements and is known with an accuracy of about 1%. Taking annual means of incoming solar radiation per unit area one finds that the

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* For a more complete discussion see: V. Bjerknes, J. Bjerknes, H. Solberg, T. Bergeron, *Physikalische Hydrodynamik*, Chap. XVI.

equator receives about 22% more, the poles 49% less than the mean for the whole earth. A considerable fraction of this incoming radiation (called the "albedo") is immediately reflected back to space. The albedo is largest in regions of high cloudiness and in the polar snow-covered regions. Unfortunately, the albedo is not known with great accuracy, primarily because the mean cloudiness in the various latitudes is insufficiently known. The error in the present estimated values is probably of the order of 10%, and hence the net amount of solar energy received in the different latitudes is known with the same degree of approximation.

The total amount of solar energy thus received by the system earth-atmosphere must be compensated by an equal amount of terrestrial radiation from this system back to space. This radiation originates partly from the earth, partly from the clouds and partly from the water vapor in the atmosphere. Without going into the rather complicated radiation process in detail, it may be stated that the approximate value of the total outgoing radiation can be calculated. The factors of uncertainty are the mean cloudiness and the temperature of the clouds, and the radiative properties of water vapor. In the last few years these properties have been investigated in the United States by extensive and successful studies where theoretical research, laboratory experiments and empirical measurements have worked hand in hand. Through these investigations, which still are in progress, reliable data for the radiation and absorption of water vapor will probably soon be available.

Several computations of the total outgoing terrestrial radiation have been carried out in the past decade, using the available data for the absorption of water vapor and the cloudiness. A check on the computation is that the total outgoing radiation from the whole hemisphere equals the total incoming radiation. The results of the different computations agree reasonably well, and may therefore be considered at least qualitatively to represent real conditions. The result may be summarized as follows:

In low latitudes the net incoming solar radiation exceeds the outgoing terrestrial radiation; in high latitudes the opposite is true. The boundary between the two regions is located at about 35° lat. Considering one hemisphere only, the total gain of energy in the equatorial region, which equals the total loss in the polar region, averages about $3 \cdot 10^{12}$ kilojoules per sec. The figures should of course be taken only as an indication of the order of magnitude of the heat amount in question.

Since the mean temperature field of the atmosphere remains unchanged, the total heat gain must be transported to the region of heat loss. Hence there must exist a heat transport of about $3 \cdot 10^{12}$ kilojoules per sec across the latitude of 35°. Similarly the heat transport across any given latitude may be computed. This heat transport is carried partly by the ocean currents, partly by the wind systems of the atmospheric circulation.

Before the heat transport carried by the atmospheric circulation can be derived, which is the final aim, the amount which is transported by the ocean must thus first be separated out. In principle the procedure to follow is clear. All that is needed is to find the heat exchange between the atmosphere and the earth. A considerable part of this exchange, which is carried by the latent heat of water vapor, can be determined from the mean values of evaporation and rainfall. One part is due to the radiation between the earth and the clouds, and finally one part is due to the radiation between the earth and the water vapor in the atmosphere. This heat exchange has not yet been determined. The principal reason is that the radiation of water vapor enters as a much more important factor in this calculation than in the determination of the total terrestrial radiation. With the present inaccurate values for water vapor absorption, any quantitative determination of the heat exchange between the atmosphere and the earth would therefore lead to completely misleading results. As soon as the above mentioned radiation data for the water vapor are available, one of the principal obstacles for a quantitative theory of the heat balance in the atmosphere will thus be removed.

In the absence of any information on the heat transport of the ocean, an attempt has been made to estimate the intensity of the meridional circulation of the atmosphere under the assumption that the total meridional heat transport is accomplished by the circulation of the troposphere. The simplest possible circulation model was chosen: The air sinks down at the pole, moves southwards along the surface of the earth towards the equator where it ascends and finally returns northwards aloft. This motion is assumed to be the same in every meridional plane, so that the troposphere of one hemisphere is considered as one single vortex ring circulation unit. This model is both dynamically impossible, and differs greatly from real conditions as far as we can judge. Nevertheless, it has been chosen, first because it is relatively simple to compute the heat transport of this model, and secondly because it will give a rough idea of the intensity of the meridional circulation required for the heat transport. The computation gives a maximum velocity of about 2.5 m/sec at 35° lat. N and decreasing velocities on either side of this latitude. The vertical velocities are of the order of 5 mm/sec. These figures have, of course, no real significance. The important thing about them is that they are found in the range we should expect, and that therefore the *method* which is applied here gives good hope of quantitative results.

Thus when the computation is repeated with more accurate values for the heat transport, derived from reliable data for the radiation of water vapor as outlined above, and when a circulation model is chosen which more closely resembles real conditions, a real quantitative determination of the intensity of the meridional circulation may be expected. As will be shown later, the generation and dissipation of kinetic energy in the atmosphere is intimately connected with the meridional circulation. Therefore, the knowledge of the intensity of this circulation is fundamental for a quantitative dynamic circulation theory.

The above brief discussion of the atmospheric heat balance has been presented to show that the application of the principle of this balance provides a method by which the intensity of the meridional component of the mean atmospheric circulation can be quantitatively determined. As far as we know, this is the only method at our disposal, and even this method is of no practical significance before more reliable radiation data are available.

Until then only a qualitative circulation model can be built up. In the next two sections it will be attempted to derive such a qualitative model. The following synthetic method of procedure will be used: First, under D below, a mean *zonal* circulation model will be derived which fits the observed mean conditions as closely as possible. Then in the following sections a certain mean *meridional* circulation will be superimposed on this zonal circulation. The meridional circulation will be introduced in several steps, guided by the general dynamical laws which must be satisfied when this circulation is started and later maintained as a steady state, under the influence of the dynamical factors which operate in the atmosphere.

D. The Mean Zonal Circulation*

The mean wind systems of the atmospheric circulation cannot be determined directly from wind observations. All direct wind measurements from the surface of the earth are greatly influenced by surface friction and local orographic conditions. Pilot balloon measurements do not suffer from this discrepancy, but since they give only information of clear weather conditions, no reliable mean values can be derived from these observations.

However, the equations of motion provide a relation between the motion and the pressure field, through which the motion may be determined when the pressure field is known. For steady state, horizontal motion, this relation becomes extremely simple: The velocity is parallel to the horizontal isobars and reciprocally proportional to the distance between consecutive unit isobars. When the pressure field is known, say by the isobars in a number of horizontal surfaces, for instance, one for each dyn. km of elevation, the steady state horizontal motion which corresponds to this pressure field is completely known, and the isobars in the horizontal surfaces represent the simplest graphical representation of the velocity field.

The mean pressure distribution at sea level is known with good approximation. This distribution shows a marked tendency to symmetry with respect to the axis of the earth, most clearly developed in the southern hemisphere, where the following characteristics are significant: A ring of moderately low pressure at the equator, the subtropical belt of high pressure at about 30° lat., a circumpolar ring of low pressure at about 65° lat., and moderately rising pressure from this latitude to the pole.

In the northern hemisphere there is considerable departure from this distribution, arising from the asymmetric distribution of the continents. When this asymmetry is smoothed out, the resulting mean zonal pressure field for the northern hemisphere will have the same main features as mentioned above.

This zonal pressure distribution is associated with the following zonal wind systems: (1) The trade wind zone, from the equator to about 30°, with east winds in the "free" atmosphere just above the layer of surface friction; (2) The west wind zone from about 30° to about 65°, with west winds in the free atmosphere; (3) The polar east wind zone from about 65° to the pole, with east winds in the free atmosphere. Due to the frictional influence in the surface layer, the wind systems are here retarded and have everywhere a systematical component towards low pressure. This effect will be considered in the next section in connection with the meridional component of the circulation.

The pressure distribution at higher levels can be determined only indirectly by means of the pressure at sea level and the mean temperature distribution with height. Statistical examination of temperature records from aerological stations in various latitudes indicates that the mean lapse rate within the troposphere does not vary appreciably with latitude. Excepting the ice-covered polar regions and the cold continents in winter, one finds on the average a temperature decrease of 5°C per km in the lower troposphere and 7°C per km in the upper troposphere. The mean temperature at higher levels can therefore be estimated with some approximation from the known surface temperatures.

Following this method the pressure distribution has been determined for the northern hemisphere at every 2 km level up to 8 km.† The most significant feature of these high level pressure maps is a large polar low which occupies the whole region north of the sub-

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* For a more complete discussion see, for instance, *Phys. Hydr.*, Chap. 15.

† Shaw: "Manual of Meteorology," Vol. II.

tropical belt of high pressure, and which becomes more and more intense with increasing height. Thus the polar region of high pressure found at sea level is a very shallow phenomenon, thermally produced by the low surface temperatures in this region. Similarly the asymmetric disturbances of the zonal pressure distribution at sea level, caused by the continents, disappear more and more with increasing height. In the upper half of the troposphere the *real* mean pressure distribution is therefore fairly symmetric, with a deep polar low in the middle, surrounded by the subtropical belt of high pressure, and beyond this a ring of slightly lower pressure at the equator.

Corresponding to this mean pressure field the wind systems aloft are as follows: Above 2 km the whole region from the subtropical belt to the pole is occupied by west winds. This "polar vortex" grows stronger with increasing elevation and has its maximum velocities at the tropopause. The region between the subtropical belt and the equator is at all levels occupied by east winds. The subtropical high pressure belt separating these two regions approaches the equator with increasing elevation, so that the region occupied by the polar vortex of west winds increases and the equatorial belt of east winds becomes narrower with increasing height. This mean zonal wind field is shown schematically in Fig. 1 which represents a meridional cross section from the equator to the pole.

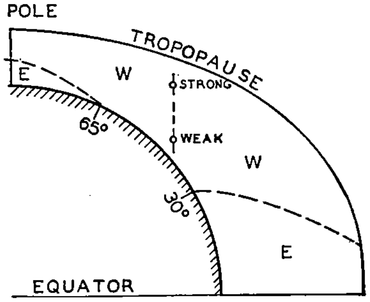

FIG. 1. Mean zonal circulation.

The mean zonal motion of the troposphere has thus the following simple characteristics: An equatorial belt of east winds extending through the whole troposphere. This belt has its greatest width at the ground, where it extends from about 30°N to 30°S, and narrows down to a width of about 10° at the tropopause. Neglecting the relatively shallow layer near the ground in polar regions, the whole remaining troposphere in both hemispheres has a motion from west to east, which increases in intensity with height.

In order to understand and visualize the dynamical properties of this zonal motion we shall consider it as "absolute" motion—i.e., with reference to a system of coordinates with the origin at the center of the earth and with fixed directions in space. The absolute motion of the earth itself is a rotation with constant angular velocity. The atmosphere rotates more slowly than the earth in the equatorial belt of east winds, and faster than the earth in the polar region of west winds. Thus the mean zonal motion of the troposphere in effect is a huge vortex whose axis coincides with the axis of the earth, and the intensity of this vortex, as indicated by the angular velocity, increases in all latitudes with increasing elevation.

Since this motion represents mean conditions, it is by definition a steady state, which can exist only if the system is completely balanced dynamically. It is easy to see, at least qualitatively, that this balance is maintained by an adjustment between the mass distribution and the velocity distribution in the atmosphere. The important characteristic of the mass distribution is that level for level we find cold and heavy masses in the polar region, warm and light masses in the equatorial region. The significant property of the velocity field is the increasing intensity of the vortex with increasing elevation, or actually with increasing distance from the equatorial plane. To realize the dynamic consequence of these two conditions, we shall for a moment let one of them operate alone, by artificially eliminating the other.

To eliminate the effect of the motion the above mentioned mass distribution is considered in a non-rotating atmosphere. Under the influence of gravity the heavier masses in the polar region would then immediately sink down and spread out southwards along the surface of the earth, with a compensating northward flow aloft of the light rising masses at the equator. This motion will soon establish horizontal stratification, and the system will perform oscillations around this equilibrium stage.

Consider next an atmosphere with horizontal stratification and rotating as a vortex whose intensity increases with the distance from the equatorial plane. This atmosphere is in complete equilibrium with respect to the force of gravity, but the centrifugal forces due to the rotation are not in equilibrium. At high levels where the vortex has its greatest intensity, the centrifugal forces are larger throughout the atmosphere than near the ground where the vortex is weaker. The result is a circulation southwards aloft, sinking at the equator and compensating flow northwards at the ground. This circulation will, of course, immediately disturb the horizontal stratification, so in the subsequent motion the retarding influence of gravity should be considered. To examine the dynamic effect of the rotation under the simplest possible conditions we assume that the horizontal stratification of the mass distribution is maintained by a suitable distribution of heat and cold sources. The gravity will then remain inactive, and it is only necessary to consider the centrifugal forces during the subsequent meridional circulation. During this circulation, assumed identical in every meridional plane, a chain of air particles lying on a parallel circle will always remain on a parallel circle. These individual fluid parallel circles will expand in the southward flow aloft, contract in the northward flow at the ground, and since no forces are acting along the circles, their angular momentum will remain unchanged. This also follows directly from the circulation theorem. The fluid parallel circles do not embrace any solenoids, therefore their circulation is constant. It is readily seen that the circulation and the angular momentum are proportional and have the ratio 2π. On the expanding fluid circles aloft the angular velocity, hence the centrifugal force decreases, on the contracting circles at the ground the centrifugal force increases, and soon a balanced stage is reached, around which the system will oscillate.

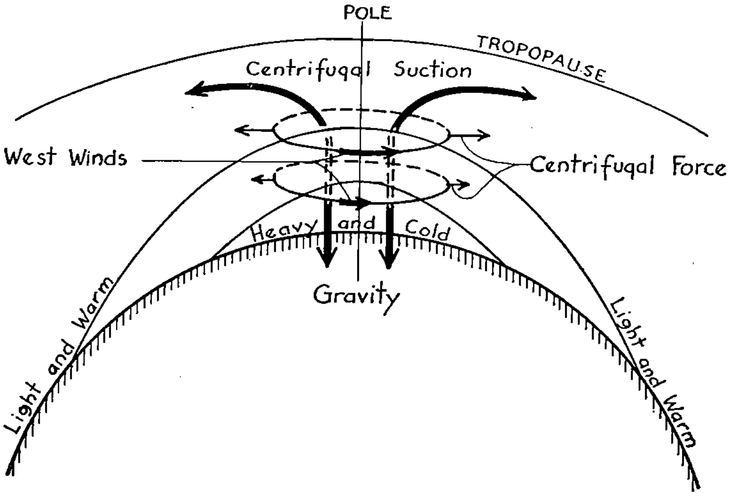

FIG. 2. Schematic balance between gravity and centrifugal suction.

When finally the two effects are operating together, the gravity tends to generate a "direct" meridional circulation around the solenoids with northward flow aloft, southward flow at the ground; the centrifugal forces due to the rotation tend to produce a "retrograde" meridional circulation, and the steady balanced state is established when these two conflicting tendencies have equal strength in every point. The dynamical balance of the atmosphere thus operates on the same principle as that of a centrifugal pump: The heavy masses near the axis are prevented from sinking down by the "suction" arising from the stronger intensity of the vortex aloft.* This dynamical balance between gravity and centrifugal suction is schematically illustrated in Fig. 2.

It may be pointed out that the dynamical principle which was the basis for the derivation of the velocity field from the empirically determined pressure field is identical with the one applied above to explain the dynamical balance. Both are direct consequences of the equation of motion, which must be the same in both cases. The only difference is that in the former case it was necessary to consider the pressure field, whose dynamical consequence is more difficult to visualize. As already mentioned, the pressure field is the one meteorological element which is directly accessible through observations. From the surface pressure the pressure is determined level for level with the aid of the hydrostatic equation. The temperature and hence the mass distribution are decisive for this pressure variation with height. When by this method the horizontal pressure distribution is determined for any given level, the steady state is characterized by the balance between the pressure gradients and the coriolis forces, from which the velocity can be determined. There is thus, on the one hand, complete correspondence between the mass distribution and the horizontal pressure gradient. On the other hand, the centrifugal forces in absolute motion correspond to the coriolis forces in relative motion. The two methods therefore describe the same dynamical balance, only in different ways. They are both useful, and supplement each other: The first method provides the simplest solution of the practical problem of deriving the velocity field from the pressure field, under steady state balanced conditions. Through the second method the dynamical nature of this balance may be most clearly visualized.

E. Dynamical Properties of the Meridional Circulation

The second method applied in the preceding section to explain the dynamic balance of the mean zonal circulation will be found even more helpful in the analysis of the dynamical properties of the meridional circulation. It has already been shown that a meridional circulation must exist in order to transport the heat excess in low latitudes to the region of heat deficit in high latitudes. With any type of meridional circulation the dynamical balance of the zonal circulation derived above cannot be complete. But if the velocity components in the meridional circulation are small compared with the zonal components, which empirical evidence indicates, the departure from dynamical balance of the zonal motion should not be expected to be large. The following discussion will therefore be based on the assumption that the zonal motion is approximately balanced, and that the nature of the small departure from dynamical balance is indicative of the sense of the meridional circulation.

In the derivation of the mean zonal motion it was assumed that this motion represents a steady state. The simplest condition under which a steady state can be maintained is that no external factors are operating on the system to disturb the state of motion. A steady state may also be imagined where several disturbing factors are operating simultaneously where the effects of the different factors tend to cancel each other. The mean circulation of the atmosphere is a system of motion of this latter type.

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* The product of the angular velocity and the square of the radius.

There are two factors continually in operation in the atmosphere which tend to disturb the dynamical balance of the zonal motion. One factor is thermal in origin and arises from the uneven distribution of the heat supply in the atmosphere. The second factor is the friction between the atmosphere and the surface of the earth, and the internal friction or eddy viscosity due to turbulence. First the dynamic nature of these two factors will be considered. Then their effect on the dynamical balance of the zonal motion will be discussed.

F. Thermal Disturbance of the Zonal Circulation

It has been already shown (p. 10 et seq.) that the equatorial region has a net gain of heat, the polar region a net loss. This fact led to the qualitative conclusion that some kind of meridional circulation must exist to maintain the heat balance. To investigate the dynamical consequence of this uneven distribution of the heat supply in its simplest form it is assumed that for a moment no meridional circulation exists and that the zonal motion is completely balanced. This zonal motion and the corresponding mass distribution are assumed to represent mean conditions as outlined in the preceding section. It is further assumed that the disturbing influence of surface friction and turbulence does not exist. If these conditions exist at a certain moment, the equatorial air masses will, during a subsequent time interval, receive a certain amount of heat, the polar air masses lose the same amount. Since, according to the above assumption, no meridional circulation exists, this thermal impulse will result in an increase of temperature of the tropical air masses, a temperature drop in the polar air masses, whereby the temperature contrast between the equator and the pole is increased. The departure of the mass distribution from horizontal stratification is therefore now greater than in the initial balance state, and the tendency of the forces of gravity to restore horizontal stratification is correspondingly stronger. The thermal impulse, on the other hand, has no direct effect on the zonal motion; therefore this motion, and hence the centrifugal forces, remain unchanged. These centrifugal forces are then no longer able to balance the increased restoring forces of gravity; hence a "direct" meridional circulation driven by the excess in gravity forces will be the result. This meridional circulation will immediately disturb both the mass distribution and the zonal circulation, and the argument therefore holds only for the initial stage when the meridional circulation begins. In order to follow the dynamical consequence of this circulation it is assumed that no new heat is added to or withdrawn from the atmosphere during this phase of the fictitious experiment, so that the subsequent variation of the mass distribution is caused by the motion only. One effect of the meridional circulation is then that the polar masses sink and the equatorial masses rise, signifying an approach towards horizontal stratification and hence a weakening of the restoring effect of gravity. The second effect is that the fluid parallel circles aloft are displaced toward the north and contract, the fluid parallel circles near the ground are displaced towards the south and expand. This corresponds to an increase in angular velocity at high levels, a decrease in angular velocity at low levels: The atmospheric "vortex" becomes weaker at the ground, stronger aloft, and the centrifugal "suction" of the vortex becomes stronger. Therefore, in brief, the result of the meridional circulation is a weakening of the effect of gravity, an intensification of the centrifugal effect, and soon a stage is reached where a new balance is established between the centrifugal forces and gravity. In the absence of any damping the atmosphere would perform oscillations around this new equilibrium position. By choosing a small initial thermal impulse and letting it be introduced slowly, these oscillations may be avoided.

To sum up the conclusions derived above: When the atmosphere initially has a balanced zonal circulation, and this circulation is disturbed by a thermal impulse of the nature assumed above, the final result of this disturbance will be a new balanced zonal motion which is reached through a *direct* circulative displacement of the air masses in meridional direction. The final balanced state is characterized by a greater rate of increase of the intensity of the atmospheric vortex with height, and a greater temperature contrast between the equator and the pole.

It is then easy to indicate the effect of a continual thermal process whereby heat is continually introduced in equatorial regions, and continually withdrawn in polar regions. In the absence of frictional damping a direct meridional circulation will be generated through which the atmosphere tends to maintain a dynamical balance of the zonal circulation. As a consequence of the meridional circulation, the intensity of the vortex becomes continually stronger at high levels, weaker at low levels. The centrifugal "suction" of the vortex increases accordingly, and the dynamical balance of the zonal motion can be maintained only by a corresponding increase in the strength of the restoring effect of gravity, which means the temperature contrast between the equator and the pole becomes greater. With increasing temperature in the equatorial region the outgoing terrestrial radiation must increase until finally a stage is reached where the outgoing radiation is equal to the incoming solar radiation. Similarly the cooling in the polar region approaches a stage of radiation balance. At this point no thermal impulse is present, the meridional circulation ceases, and the atmosphere has a final state of balanced zonal motion, which actually would be approached asymptotically. The analysis of the thermal disturbance has here been carried to its final consequence if it was permitted to operate alone. The result has, of course, no practical significance, since the damping effect of turbulence and surface friction also operates in the atmosphere and prevents the growth of the zonal circulation. The effect of this factor on the dynamical balance of the zonal motion will now be considered.

G. Integral Effect of Friction on the Zonal Circulation

The general effect of friction is to even out all relative velocity differences. Regardless of what the initial state of motion in the atmosphere is, it is therefore easy to predict the final state if friction, and this factor only, is permitted to operate on the system sufficiently long. Every velocity difference between the earth and the surface layers will be eliminated by surface friction, and all velocity differences within the atmosphere will be smoothed out by the eddy viscosity, so that in the final state the atmosphere and the earth will rotate as a solid system with the angular velocity of the earth.

To investigate the dynamical action of friction in more detail the atmosphere will be assumed to have the same initial motion as in the preceding case when the thermal effect was examined—i.e., no meridional circulation, and a dynamically balanced zonal circulation corresponding to mean conditions. For the moment it is assumed that no thermal action operates, so that the friction is the only factor which tends to modify the motion. The final state is then, as already mentioned, solid rotation, which is characterized by horizontal mass distribution. In the absence of any external thermal action the redistribution of the masses from the initial to the final state must have been brought about by the motion; hence it may be concluded that there has been a net meridional circulative displacement with sinking and flow towards the south of the cold masses in polar regions and a compensating northward flow aloft of the warm tropical air masses. The underlying dynamical reason for this meridional circulation is, of course, that the centrifugal suction of the atmospheric vortex becomes weaker as the eddy viscosity tends to diminish the velocity increase with height, so that the dynamical balance between the centrifugal forces and the forces of gravity is disturbed. The latter dominates, whereby a direct meridional circulation tends to be generated. The comparison between the initial and the final state therefore indicates that if friction alone operates on the mean zonal circulation of the atmosphere, certain meridional circulations are generated, and the net result is a direct circulative displacement. However, it is impossible by this simple method to find out anything about the details of the actual meridional motion. For this purpose it is necessary to study the instantaneous dynamical action of the friction.

H. Dynamical Nature of Atmospheric Friction

The dynamical process underlying and giving rise to the atmospheric friction is in the nature of a transport of momentum directed from regions of high towards regions of low values of the momentum. This momentum transport is maintained by the irregular fluctuations or eddies of the turbulent motion of the atmosphere. The dynamical nature of these eddies, and hence of the momentum transfer, is not yet completely known, and it would lead too far afield to enter upon a discussion of this very complicated subject. Only some general remarks will be made to clarify the main aspects of the problem.

Even without knowing the details of turbulent motion it is obvious that the intensity of the momentum transport must depend upon the momentum gradient, the scale of the eddies, and the period of the fluctuations. From this it may be concluded that the region of the atmosphere where the momentum transfer, hence the friction, is most active is the layer next to the surface of the earth, and for two reasons:—On the one hand the surface wind is greatly retarded as compared with the wind velocity in the free atmosphere, giving rise to a steep vertical velocity gradient. On the other hand, the roughness of the surface of the earth is a very effective agency for the creation of turbulent eddies of high frequency. This results in a strong downward directed flow of momentum through the surface layer.

At high levels the velocity variation is much more gradual both in meridional and vertical direction; hence even if turbulence is present the resulting transport of momentum and therefore the friction must be much weaker in each locality than in the surface layer. Nevertheless, the *total* effect of this friction upon the motion of the atmosphere as a whole may be as important as the surface friction, since it operates through a much thicker layer. Up to very recently it was assumed that the motion at high levels was approximately laminar, and that therefore the friction could be neglected here, although no observational evidence in support of this assumption had been obtained. On the contrary, the diffusion of smoke and volcanic dust and the scattering of small balloons indicate that turbulence always is present at all levels. The necessity of considering the high level turbulence in the dynamical study of large-scale atmospheric motion was first realized by C.-G. Rossby during his theoretical studies connected with the present research project at M.I.T., and the results he has obtained from the subsequent investigation of this turbulence have already led to a much clearer understanding of the dynamical control of the large-scale motion in the atmosphere.

The dynamical nature of the high level turbulence is not clear yet, but some general conclusions may be drawn on purely theoretical grounds. The flow at high levels is not disturbed by an external mechanical agency, as was the case in the surface layer where the roughness of the ground operates. Therefore the high level turbulence must arise spontaneously, due to some instability of the general flow. The stability of the flow is determined by the following three factors: (1) Hydrostatic stability controlled by the gravity and determined by the stratification (temperature lapse rate); (2) Inertia stability due to the rotation, controlled by the centrifugal forces and determined by the variation of the angular velocity normal to the axis of the earth; (3) Shear instability. For large-scale atmospheric flow the first and second factors are always found to be stabilizing. The third factor is always destabilizing whenever the flow has any shear, and the instability is proportional to the velocity gradient. There is thus in the vertical direction a conflict between shear instability and hydrostatic stability, the latter generally being much stronger. The conditions are therefore rather unfavorable for the development of any appreciable vertical turbulence. In the horizontal direction, on the other hand, the destabilizing effect of the shear is counteracted only by the inertia stability. It has not yet been possible to obtain an exact solution for a system where these two factors operate together. However, by examining the two factors separately it may be shown qualitatively that the shear dominates for perturbations with small "wave-length," the inertia stability dominates for long wave lengths. The critical wave-length depends upon the strength of the shear, and increases with the shear. But further theoretical study is needed to clarify this problem, and to find a quantitative expression for the critical wave-length.

Thus theoretical considerations also indicate that the instability necessary for the generation of high level turbulence always is present in the atmosphere, and that this turbulence must consist of quasi-horizontal eddies. The scale of these eddies, or rather the upper limit for the scale, is determined by the strength of the horizontal shear and increases with increasing shear.

I. The Empirical Method of Isentropic Analysis

Rossby felt that if large-scale high level eddies are a frequent development in the atmosphere, it should be possible to establish their presence through observations, if the meshes in the aerological network are smaller than the dimensions of the eddies. The network over the United States is closer and covers a larger area than anywhere else in the world, and should therefore be most favorable for a successful search for these eddies. At first sight it should seem quite easy to detect large-scale eddies directly from pilot balloon observations. This simple method must be abandoned for three reasons: It can be used only over regions with clear weather; there is no way of identifying a disturbance in the flow from one day to the next; and most important of all, it is impossible to determine whether the disturbance has been generated by thermal action or has developed spontaneously due to instability of the flow.

However, the eddies will also reveal their presence indirectly through the accompanying modifications of the pressure field and the mass distribution, and should therefore be detectable through an appropriate analysis of these fields. For this purpose a special technique is required, by means of which the effect of the eddies may be separated from the effect of large-scale variations in the flow, and which makes the identification of individual air masses from day to day possible. For this second requirement the analysis must concentrate on the study of conservative or quasi-conservative properties of the air. No property of the air is completely conservative, but at high levels the potential temperature and the specific humidity are approximately conservative for short time intervals. (Regions where condensation occur are, of course, excepted from this rule.) A short while ago Rossby defined a third quasi-conservative quantity called "potential isentropic vorticity." These three properties are independent of each other. Therefore, by determining their three-dimensional distribution day by day, the problem of finding the trajectories of the individual air particles is solved in principle.

For the practical solution of the problem a special technique, called "isentropic analysis," has been developed at M.I.T. With the aid of this method, which has been described and illustrated in a previous report, the existence of large-scale horizontal eddies at high levels had been empirically verified. Without going into details, it may be stated that these eddies show a good correlation with the horizontal velocity gradients of the zonal circulation, in agreement with the qualitative theoretical conclusion derived above. The eddies are located in regions where the horizontal shear is appreciable and are particularly well-developed and have large dimensions during periods with abnormally strong shear. The observed eddies have dimensions ranging between 500 and 1000 km. These probably correspond to the "critical wave-length." Smaller eddies which should be even more favored by the dynamical instability of the flow and which undoubtedly exist, escape our notice in the relatively open network of aerological stations.

To sum up this rather lengthy discussion of the dynamical nature of the friction in the atmosphere, the following main features are important for the present discussion: (1) The internal friction or eddy viscosity in the atmosphere is the result of a transport of momentum through the action of atmospheric turbulence. (2) In the surface layer the momentum transport is large and mainly vertical, due to the large vertical momentum gradient, and the production of intensive vertical turbulence by the mechanical action of the rough surface of the earth. This results in strong friction in the surface layer. (3) At high levels the flow of momentum is weak in any locality and is mainly horizontal. This is due to the fact that the shearing instability of the flow always is suppressed in the vertical direction by the much stronger hydrostatic stability. In horizontal direction the shearing instability will dominate the stabilizing influences of the rotation for perturbations below a certain critical wave length. Hence high level turbulence has the character of horizontal eddies, the scale of which is determined by the strength of the horizontal shear. Isentropic analysis of aerological data over the United States has given empirical evidence for the presence of large-scale high level eddies, ranging in size from 500 to 1000 km. With this factual knowledge it is possible to return to the problem of the dynamical effect of friction on the balanced zonal circulation.

J. Frictional Disturbance of the Zonal Circulation

The system under consideration is, as in the case of a thermal disturbance (see section F), a completely balanced zonal circulation representing mean conditions, and hence initially having no meridional circulation. The system will be assumed thermally isolated, so that friction is the only disturbing factor. To simplify the problem still further, friction at high levels may, for the moment, be neglected, so that only the action of surface friction will be considered. As in the case of the thermal disturbance, let the surface friction operate on the system during a short time interval. The air masses within the surface layer, which initially are assumed to move freely with velocities of the "free atmosphere" (above the surface layer), are then retarded relative to the earth.

In absolute motion this means that the surface layers of east wind (the trade wind belt and the polar region north of 65°) have a weaker east wind than the free atmosphere, hence rotate more rapidly than they did initially. The rotation of the vortex increases less rapidly with height and its centrifugal suction has diminished. The dynamical balance of the zonal circulation is disturbed with the effect of gravity dominating. The result is a *direct* meridional circulation. This circulation is, of course, confined to the atmosphere above the regions of east winds. Hence there must be two such circulation units, one in the trade wind belt and one over the polar region. The southward branches of these circulation units have relatively strong intensity and are confined to the surface layer. The compensating northward branches extend through the whole remaining part of the troposphere and are extremely weak.

The west winds at the surface are also retarded by the friction, hence the absolute rotation of these layers decreases, with a subsequent intensification of the centrifugal suction. The result is a *retrograde* meridional circulation of this part of the troposphere. The three circulation units generated by a frictional disturbance are schematically illustrated in Fig. 3.

If the surface friction is only operating through a short time interval, the subsequent meridional circulation would soon stop. The final state is a new balanced zonal circulation, with weaker zonal velocities than in the initial state. Through continual action of the surface friction the three circulation units would continue with the sense as indicated by the above discussion. Through the continual loss of momentum in the surface layer the zonal circulation would slow down until the final state of solid rotation was reached.

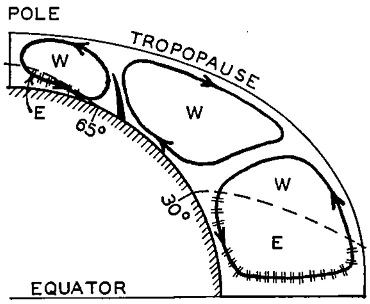

Fig. 3. Combined meridional-zonal circulation.

An important consequence of the frictionally driven meridional circulations is their effect on the mass distribution. On the borderline between the direct trade wind circulation and the retrograde west wind circulation there is horizontal divergence in the surface layer and a slow subsidence in the free atmosphere. In this region therefore the north-south temperature gradient is decreased in the lower part of the troposphere. The stratification becomes quasi-horizontal, and the increase of rotation with height required for dynamical balance is only very slight. In the upper part of the troposphere the conditions are reversed, with a steep increase of rotation with height, which seems to agree with actual conditions.

At the line of demarcation between the west wind circulation and the direct circulation of the polar east winds there is horizontal convergence in the surface layer giving rise to a sharp horizontal temperature contrast. The dynamical balance of the zonal circulation requires a strong increase of rotation with height at this point. The northern part of the west wind belt which at high levels extends further north and overruns the cold wedge of east winds should therefore be expected to be a region of very strong west winds. This is also in good agreement with general experience.

Thus through the indirect action of surface friction an important modification is produced in the west wind circulation at high levels, resulting in an intensification of the west winds to the north over the "polar front" and a weakening in the west winds to the south over the subtropical high. This signifies a relatively steep horizontal velocity gradient between these two latitudes. This is therefore the region of the atmosphere where the high level friction might be expected to be most active. Experience also shows that this is the region where the high level horizontal eddies are generally located. Through the action of this high level turbulence there is a southward transport of westward momentum: The rotation aloft thus tends to increase, thereby increasing the centrifugal suction in the region. Thus the high level friction aloft supports and helps to maintain the retrograde circulation of the west winds which is generated by the surface friction.

K. The Mean General Circulation

After having obtained a qualitative estimate of the dynamical effect of thermal and frictional disturbances when they operate separately, an attempt will now be made to estimate their joint action when they operate together. During this discussion it is useful to keep in mind the characteristic difference between the two factors and the modification of the circulation they tend to produce.

The discussion of the balance of the two factors is largely based on the principle that the total amount of kinetic energy and of potential energy in the atmosphere must remain unchanged when the circulation is maintained at a steady state. The transformation of energy associated with the thermal and the frictional action will therefore be considered separately first.

Though the thermal action the temperature contrast between the equator and the pole is steepened, thereby increasing the potential energy of the system. One part of this potential energy is released through the subsequent meridional circulation displacement and converted into kinetic energy. A thermal impulse therefore results in a direct meridional circulative displacement and an increase in both the potential and the kinetic energy.

The frictional action, on the other hand, results in three meridional circulation units, one retrograde above the west wind belt with the direct trade wind circulation to the south and the direct circulation of the polar east wind region to the north. The energy transformations are different in these units.

In the direct circulation units the immediate action of friction is to diminish the kinetic energy of the surface layer. Through the subsequent direct circulation also the potential energy of the system is diminished. This loss in potential energy must be converted into kinetic energy which in part, but not completely, will make up for the frictional loss. The net result of the frictional action in the direct circulation units of the east winds is therefore a direct circulative displacement associated with a decrease in both the potential and the kinetic energy of these units.

In the west wind belt the immediate action of friction is also here to diminish the kinetic energy in the surface layer. Through the subsequent retrograde displacement the air masses are forced to rise to the north and sink to the south, thereby increasing the potential energy of the system. The net result here is therefore a gain in potential energy and a relatively large loss in kenetic energy, partly through friction and partly used to build up the potential energy.

When the general case is now considered, where the thermal and frictional factors are operating together, the first question to be answered is what type of meridional circulation will finally be chosen. In the regions of east winds the answer is easy. Both thermal and frictional action tend to generate direct circulation; therefore also their joint action must result in a direct circulation. In the west wind belt the two effects tend to produce opposite circulation. An indication of which of these tendencies will dominate may be obtained by considering the frictional effect in the surface layer. No thermal action can prevent the frictionally caused poleward flow in this layer as long as the surface winds are from the west. The surface layer at least must therefore represent a branch of a retrograde circulation unit. There is, of course, the theoretical possibility that there may be a shallow retrograde circulation in the lower layers of the west wind belt and a direct circulation driven by thermal action above it. In fact, older circulation theories were based on this assumption. However, by considering the high level friction Rossby* has recently shown that one single retrograde circulation unit of the west wind belt represents a system whose dynamical properties can be simply explained and as far as can be judged, is in accordance with observational facts. This retrograde circulation of the whole west wind belt must therefore be accepted as the most probable hypothesis. It is clear that the dynamical properties of the direct and retrograde meridional circulations are different; each of them therefore requires a separate consideration.

The dynamical properties of the direct circulation units of the east winds are almost self-evident. Through thermal action both the potential and kinetic energy of the system increase, through frictional action both decrease. When the effects have equal magnitudes a steady state is established corresponding to a certain approximately balanced zonal circulation, and superimposed on this a direct meridional circulation with a narrow and relatively intense southward branch in the lower layers, a deep northward moving branch extending through the rest of the troposphere. This steady circulation is also stable. If the thermal action is increased above the mean value, say by a variation in the incoming radiation, this will result in an increase in the potential energy, with a subsequent increase in the meridional circulation and in turn an increase of the zonal circulation. The frictional action is thereby increased, and soon a new balanced steady state is reached, characterized by a somewhat more intensive circulation, and a steeper horizontal temperature gradient. The seasonal variation in the incoming radiation produces just such a variation in the intensity of the circulation from a maximum in winter when the thermal action is large to a minimum in summer when the thermal action is small. A similar variation in the circulation is caused by a decrease of the frictional action. Also then the thermal action would dominate until a new steady state is established with stronger zonal circulation and a stronger horizontal temperature contrast. Empirical evidence of this effect is found in the difference between the northern and southern hemisphere. The almost completely ocean-covered southern hemisphere has a much smaller frictional action and hence a stronger zonal circulation and a larger temperature contrast between the equator and the pole.

The dynamical mechanism of the retrograde west wind circulation is more complicated. Through the frictional action in the surface layer, which is the principal control of this circulation, there is a tendency to a slow increase in the potential energy and a relatively large loss in kinetic energy. Through a *normal* thermal action both potential and kinetic energy would tend to increase. Since the net result of both actions is a retrograde

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* Transactions of the American Geophysical Union, 19th Annual Meeting, 1938.

displacement, there is a net loss of kinetic energy and a net gain in potential energy. Thus no steady state can be established under the influence of these two factors only. There exist, however, also both a secondary thermal action and a secondary frictional action whose effect has not yet been considered.

The secondary thermal action is a consequence of the latent heat of water vapor. Through the process of evaporation the surface layers of the atmosphere withdraw heat from the surface of the earth which does not result in any temperature rise of the air. This latent heat is transported with the humid surface currents and is released in the ascending branches of the circulations where condensation takes place. Applied to the retrograde circulation the thermal action of the latent heat causes a heat loss in the south (from the earth) and a heat gain to the north, thereby tending to diminish the potential energy of the system.

The secondary frictional effect is a consequence of the high level turbulence referred to above, which has its maximum intensity in the west wind belt. The large-scale quasi-horizontal eddies of this high level turbulence transport momentum southward from the region of stronger west winds to the north. The west winds here are maintained by a similar frictional coupling with the west winds of the direct polar circulation unit further north. The high level turbulence thus provides a mechanism by means of which momentum, and hence kinetic energy, is fed into the retrograde circulation unit from the direct circulation unit to the north. This effect alone therefore causes an increase of the kinetic energy of the retrograde west wind circulation. It is also important to note the direct dynamical consequence of this effect. The kinetic energy is added at high levels, thereby increasing the centrifugal suction of the vortex which helps to maintain the retrograde circulation.

The dynamical process of this circulation is determined by the four factors discussed above. The combined effect of the two primary factors is a tendency to increase the potential energy and to decrease the kinetic energy. The secondary thermal effect of the latent heat causes a decrease of the potential energy, and the secondary effect of the high level turbulence an increase of the kinetic energy. The steady state is reached when these four effects are balanced.

The dynamics of the whole mean meridional circulation of the troposphere can now be summed up. The two direct circulations are driven directly by thermal action of the uneven supply of solar energy. The kinetic energy created through the circulation of the polar unit is partly lost by surface friction, and partly fed into the retrograde circulation of the west wind belt. The west wind circulation in turn consumes this kinetic energy in part through the friction in the surface layer and in part in making up for eventual loss in potential energy, notably due to the latent heat. There is, therefore, an important dynamical coupling between the polar unit and the retrograde unit of the west wind belt, and the steady state also requires complete balance between the two units. The third circulation unit of the trade winds is in our analysis an independent unit balanced in itself. It is conceivable, however, that also some type of coupling exists between the trade wind circulation unit and the circulation unit of west winds, so that this circulation is driven from both sides, and thus the circulation of the whole troposphere is dynamically coupled into one single dynamical system. In case such a coupling between the trade wind and the west wind circulations exists, it would probably be strongest near the tropopause where a strong horizontal velocity gradient exists in this region, but there is as yet no clear evidence in support of this hypothesis.

An interesting question is whether the circulation model derived here is the only possible circulation of the troposphere. The discussion presented here does not answer this question since the model has been derived from the empirically determined zonal circulation, thus an initially prescribed velocity distribution. The question is then whether the final steady state is independent of the initial velocity distribution as long as the thermal and frictional action are the same. Several attempts have been made to answer this question. A recent and very convincing analysis has been carried out by Rossby. He assumes the atmosphere to be initially at rest, and shows qualitatively that the final steady state must have the three circulation units which have been derived here. Rossby's discussion will be published in an article in the U. S. Dept. of Agriculture Year Book for 1941.

L. Zonal Perturbations of the General Circulation

In the preceding discussion it was assumed that complete zonal symmetry always exists, so that the motion is identical in every meridional plane. If the surface of the earth was uniform in every respect this might approximately be the case, which is also suggested by the conditions in the southern hemisphere. It was already pointed out at the outset that the circulation here shows a remarkable symmetry with respect to the axis of the earth.

In the northern hemisphere the distribution of continents and oceans is responsible for a radical departure from zonal flow in the lower part of the troposphere. This is due to the difference in thermal properties of land and sea, and to the difference in roughness, thus causing abrupt changes in both thermal and frictional factors along the coast lines. The direct effect of this asymmetry in the thermal and frictional actions is the so-called monsoon circulation systems. It is beyond the scope of this general discussion to give a description of these systems and their dynamical control.

Another indirect effect of the thermal and frictional perturbations concentrated along the coast lines was discovered last year by Rossby.* He has shown that these perturbations generate a stationary wave in the zonal currents of the general circulation where these currents are crossing coast lines running in a north–south direction. The wave length of these waves which remains bound to the coast line is proportional to the mean velocity of the zonal current. Through these stationary waves the zonal circulation breaks up into closed horizontal circulation units, the so-called "main centers of action" whose position and intensity are of decisive importance for the development of the weather.

It is impossible in this report to enter upon a discussion of the dynamical explanation of these stationary waves in the zonal circulation, and finally how the horizontal circulation units thus generated are supplementing the model of the idealized combined zonal-meridional circulation. It must suffice to say in conclusion that the dynamical action of this final model is now known and understood in its main features, and that it also is in good accord with observational evidence to the extent it can be checked. But much effort is still required in the fields of synoptic meteorology, aerology and radiation before the circulation process of the atmosphere can be determined quantitatively.

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* This work is as yet unpublished.

SECTION II. STATISTICAL AND SYNOPTIC RELATIONSHIPS

In the following discussion of possible synoptic and statistical relationships in the general circulation of the atmosphere which have been selected for investigation, an effort has been made as far as possible to arrange the topics treated in such a manner that the discussion progresses from the large-scale more general phenomena to the more specific local phenomena. Thus the discussion progresses from concepts involving the general circulation as a whole, and its correlation with local anomalies, to concepts involving the different branches of the general circulation and their relationship with local anomalies, particularly in North America.

A. The Zonal Index, Its Fluctuations, and Its Resolution into Latitudinal and Longitudinal Zones

1. *Basis for computing the zonal index. Momentum in the westerlies*

In the last section reference was made to the belt of west winds which exists in the mean on the Northern Hemisphere. The existence of the westerlies is well shown by a profile of the average zonal pressure distribution, wherein average pressure is plotted against latitude. In Fig. 4 is shown the mean profile for the winter of 1938–39, November through April, based on pressures taken at every 5° latitude by 10° longitude intersection from the daily Northern Hemisphere charts. Pressure is plotted against the sine of the latitude so as to make area under the profile proportional to total mass of air. Where the profile slopes down toward the right a net east wind is indicated, and where it slopes down toward the left a net west wind is indicated. The slope at any point is proportional to the mean of the east and west components of the wind velocity at that latitude. A normal profile for November to April, based on mean monthly pressure maps given in Shaw's Manual of Meteorology, shows that the principal band of westerlies lies between about 35° and 55° latitude. It was on the basis of this pressure profile that the index of the strength of the westerlies was computed. Since the actual velocity is of less importance than the changes in velocity from week to week, the index of the average velocity is taken simply as the pressure difference between 35° and 55° latitude. This pressure difference should be approximately proportional to the average velocity if latitudinal differences in density are neglected and the pressure on the profile varies as the cosine of the latitude.

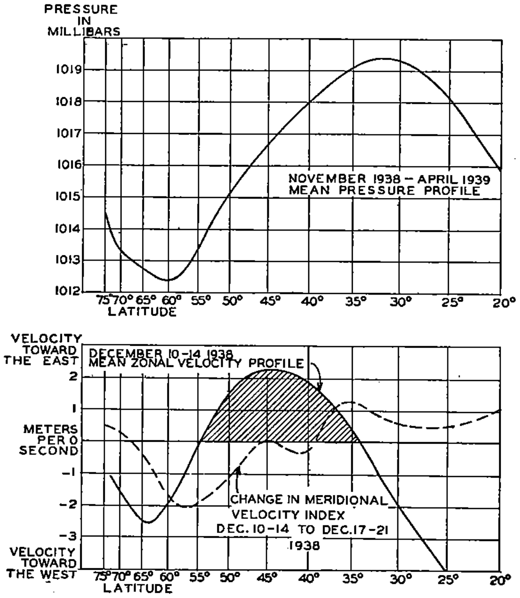

Fig. 4. Northern Hemisphere mean pressure and velocity profiles.

In order to test how well the pressure difference between two fixed latitudes represents the total momentum in the westerlies, the pressure gradients for each 5° latitude band were multiplied by the appropriate factor to obtain the mean geostrophic wind velocity, account being taken of the normal variation in density with latitude due to the variation of temperature. These mean wind velocities were then plotted on a sine latitude scale to obtain a velocity profile for each week. A typical weekly mean profile is shown in Fig. 4. The area between the curve and the axis of zero velocity represents total momentum of the geostrophic wind. These areas for the westerlies were measured with a planimeter to obtain a series of weekly momentum values for comparison with the zonal pressure differences. The correlation for the 33 weeks, October 15, 1938, to May 31, 1939, was 0.94. Thus the 35°–55° pressure difference appears to be as good an index of the velocity of the westerlies as can be obtained from surface pressures. Figure 7 (p. 36) shows a graph of zonal index values for three winters.

2. *Persistence in the zonal index*

As a preliminary step in finding possible ways of forecasting the index for a week in advance, a study was made of the serial correlation in the weekly index series. Values of the index are available for a total of over 240 weeks (either weekly or 5-day averages). These were divided into summer and winter series and the serial correlation found for each summer and each winter separately, also for all summer values and all winter values. The summer series included the months April to September for the years 1934–1939 and the winter series October to March for the years 1932–33, 1935–39. The serial correlation was small for all individual summer series, and for the six summer periods combined (118 weeks) the correlation was 0.23. A low correlation would have been expected, for the westerlies are known to be weak and to fluctuate irregularly in summer. The serial correlation for winter, however, is considerably higher. For the five individual winters the coefficients range from 0.38 to 0.58, with the correlation for all winters (125 weeks) equal to 0.47.

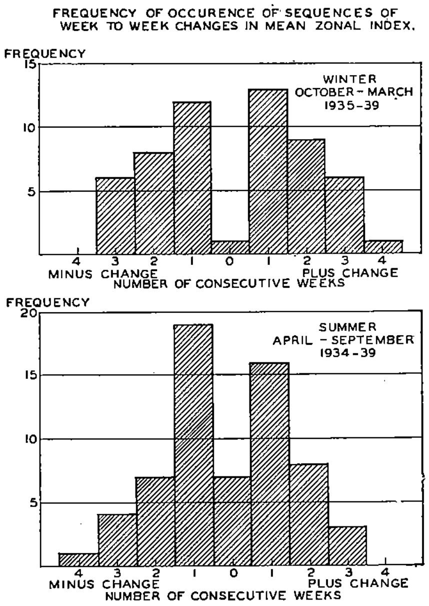

Fig. 5. Frequency of occurrence of sequences of week-to-week changes in the mean zonal index.

From the standpoint of forecasting the index it is even more important to have a measure of the persistence in week-to-week changes than to have a measure of the persistence in departures. Fig. 5 shows the frequency of occurrence of the given sequences of changes in the weekly mean index. For example, out of 100 week-to-week changes during the winter, there were 13 cases when the index increased from one week to another, with falls both before and afterward; there were 9 cases when the index increased for two weeks in succession, with falls occurring both before and afterward; etc. The frequency indicated for zero is the number of zero changes. For summer, a change of either sign is much more likely to last only one week than to last two weeks, or three weeks. In winter, sequences of two or three weeks with changes of the same sign are relatively more common and account for 70% of the changes.

It might be mentioned here that during October, 1939 to March, 1940, a persistence method of forecasting the change in the index from week to week was used, which was fairly successful. The hemisphere index was not available, but the mean pressure difference from 60°W to 160°W was used as an estimate of the zonal index, and was averaged for the week Friday through Thursday. On each Friday an average index was obtained also for the two days Thursday and Friday. The sign of the difference between this average and the average for the preceding week was used as an estimate of the change from the preceding week to the coming week. Thus if the Thursday–Friday average was lower than the index for the week just ended, a forecast was made that the index would be lower for the week just beginning than for the week just ended. The estimated change was correct in sign 18 times out of 23 weeks, or 78%. Slightly greater than 50% verification would be expected even from random data due to the overlapping of the 7-day means with the 2-day means, and due to the fact that these are change forecasts and not departure forecasts.

An estimate of the extent to which the 2-day mean may be used to forecast index magnitudes is shown by a regression of the 2-day mean on the following 7-day mean for the period given above. The correlation coefficient was 0.72, which is significantly greater than the correlation of 0.27 that would be expected if there existed no serial correlation. The standard error of estimate from the straight-line regression was 5.0 millibars, whereas the standard deviation of the 7-day means was 6.9 millibars.

Thus the forecast of the index on such a persistence basis is significantly better than chance, and when applied to the index for the whole hemisphere, should give even better results, for the reason that the serial correlation is in general higher for the hemisphere index than for the sectional index. For the usual forecast, however, the first day of the forecast period would not be included in the short-time mean used as a tendency, and the correlation of forecast and observed means would be somewhat reduced on this account.

Although it is possible, then, to make a better than chance forecast based on persistence, it is, of course, desirable to find some method based on physical rather than purely statistical grounds.

3. *Periodicities in the zonal index*

From the charts showing the trend of the zonal index from week to week it has appeared that the index during the winter tends to reach a maximum at intervals of from 5 to 10 weeks. This period between maxima is not constant, even during a single winter, nevertheless it is of interest to have some measure of the tendency toward periodic variation. Perhaps the best method of measuring this tendency is to fit cosine curves to the index data, for it is then possible to test the reality of the suspected periods by the methods of analysis of variance. Accordingly, a Schuster periodogram analysis was made of the weekly mean indices for the six winter months of the years 1932–33, and 1935–39. Periods of 5, 6, 7, 8 and 9 weeks were tried, and in addition 10 and 11 weeks for the winter 1937–38. For the winter of 1938–39, consecutive 5-day means were used instead of weekly means.

The results are shown in the table below, where R is the amplitude of the cosine curve of the given period, and σ is the standard deviation of the zonal index values. $R/\sigma\sqrt{2}$

is the correlation between the cosine curve and the zonal index. The given period is the one having the largest amplitude, but in 1937–38 periods of 7, 9 and 11 weeks were almost as significant, and had amplitudes greater than 3.2 millibars.

TABLE I

Year	Period Weeks	R Millibars	Number of Weeks Used	$R/\sqrt{2}\,\bar{\sigma}$
1932–33	5	2.7	25	.43
1935–36	8	2.3	24	.46
1936–37	8	3.1	24	.63
1937–38	6	3.8	24	.65
1938–39	45 days	2.4	32	.58

Correlation coefficients are given here only as an estimate of the proportion of the variation in the index series that is associated with the cosine curve, and not as an indication of the significance of the periodicities. The difficulties in testing the significance of periodicities have been discussed by Page* and others, who point out that the number of cases being tested is equal to the number of cycles, and not to the number of observations. It is not suggested here, then, that the index necessarily is controlled by a real periodic force of the given period length. It should be noted from these results only that there is a tendency for the principal fluctuations of the index in winter to occur in long swings occupying roughly six to eight weeks each.

4. *Sunspots and the general circulation index*

A further indication of the behavior of the principal zone of westerlies came rather unexpectedly from correlations of the zonal index with sunspot numbers. There has long been given credence by many meteorologists to statistical indications that periods of maximum sunspot activity are accompanied by increased storminess (intensified general circulation) and a southward displacement of maximum storm activity. If this is a fact, then there should be some correlation between the zonal index and sunspot activity. To check on this, annual values of the sunspot numbers (A. Wolfer) were correlated with the annual values of the zonal index for the ten-year period 1921–1930, as determined from ten years of seasonal mean pressure maps of the northern hemisphere prepared at the Weather Bureau. The zonal index first used was that given by the mean pressure difference between latitudes 40° and 60°N, because for this period it was found to average slightly larger than that between 35° and 55°N.

The correlation between these ten pairs of annual mean values was found to be −0.33. This could mean either that the general circulation was weaker during sunspot maxima, or that there occurred a latitudinal shift of the westerlies such that a smaller portion of the poleward pressure gradient lay between 40° and 60°N. If a southward shift occurred, as has been believed, with sunspot maximum, that should be shown by an increased zonal index between 35° and 55°N. The correlation of this index with the sunspot numbers for the same ten-year period was found to be +0.51. This indicates that probably the entire decrease of the 40°–60° index with sunspots was due to a southward shift of the westerlies, and that this shift was accompanied by a moderate increase in the intensity of the zonal circulation.

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* *Monthly Weather Review*, Supplement No. 39, 1940.

The correlation of the two sets of zonal indices during this ten-year period was found to be 0.478. This is surprisingly small, in view of the fact that the two zones have three-fourths of their respective pressure profiles in common. These correlations show conclusively that during this ten-year period the principal changes of the annual mean state of the zonal circulation were not in its intensity, or the steepness of the pressure profile in middle latitudes, but rather in the latitudinal shift of its maximum intensity. The facts indicated by these correlations are significant for long range forecasting if they are found to hold over long periods. However, in view of the lack of permanence manifested in the past by most correlations and relationships involving sunspots, they cannot be accepted as significant until given a much more extended check.

There is one earlier statistical observation that tends to confirm the relation suggested above between sunspots and the state of the general circulation. Sir Gilbert Walker* correlated his so-called North Atlantic Oscillation (a measure of the state of the General Circulation over the North Atlantic quadrant) with sunspot numbers. He was disappointed to find a negative correlation (he had expected a maximum of storminess with high sunspots), and he dropped that investigation forthwith. However, it will be found on checking the indices by which he measured his North Atlantic Oscillation,† that whereas all six terms contribute positively to his total index in case of an intensification of the Icelandic low and of the Azores high (i.e., an increased zonal circulation), five out of six of his terms react negatively to a southward displacement of such a pressure system. Probably, then, the negative correlation between sunspots and his North Atlantic oscillation exactly parallels the present negative correlation between sunspots and the 40°–60° zonal index. Although he was dealing with only one quadrant of the northern hemisphere, it is the one quadrant whose zonal index shows the highest correlation with that of the northern hemisphere. (See Table II, p. 33.)

5. *Relationships between the zonal index and weekly shifts of maximum and minimum points on the northern hemisphere mean pressure profiles*

The principal zone of westerlies in middle latitudes appears in the mean pressure profiles as a region of maximum poleward slope of the profile, usually bounded on the south side by the maximum point of the subtropical belt of high pressure, and on the north side by the minimum point of the polar or sub-polar trough of low pressure (Fig. 4, p. 26). It is of practical importance to inquire as to whether there is evidenced any systematic behavior in the changes of the five-day mean pressure profiles from one week to the next. This means specifically in this case as to whether there is any relation between changes in the intensity of the westerlies and their latitudinal shifts, and as to the degree to which the latitudinal displacements of the max and min points on the profiles parallel each other.

There are two lines of reasoning from which one may draw conclusions as to the nature of the changes to be expected from week to week on the mean pressure profiles. By one of these the zone of westerly winds is considered as being made up of a number of rotating annular rings extending around the earth's axis. On the assumption of a tendency to the conservation of angular momentum of each of these rotating annular rings, and allowing for the variation of the Coriolis force with latitude on the rotating earth, it

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* *Memoirs of the Royal Meteorological Society*, Vol. 4, No. 36, 1932.

† *Memoirs of the Royal Meteorological Society*, Vol. 4, No. 39, 1937, p. 120.

can be shown that a steepening of the pressure profile (increased intensity of the westerlies) should be accompanied by a poleward displacement of the zone of westerlies. This must be accompanied by a poleward displacement of the max and min points on the pressure profile. A flattening of the pressure profile should be accompanied by corresponding southward shifts. Since the energy source of such changes in intensity of the westerlies is probably to be looked for on the side of the poleward belt of cyclonic vortices, it should follow that changes in the zonal index originate on the poleward side of the westerlies, and the latitudinal shift of the min point on the profile should normally precede that of the max point.

The second line of reasoning is based on the kinematical analysis of the field of motion in which, on the surface of a rotating sphere, there exists a circumpolar ring of cyclonic vortices, and on the equatorward side of this a ring of anticyclonic vortices. These vortices may partake of uniform circumpolar motion in their respective rings. Such an ideal system corresponds in a general way to the ring of subpolar lows and that of subtropical highs which produce the min and max points, respectively, of the mean pressure profile of the northern hemisphere. The kinematical analysis of such a system indicates that a poleward contraction or equatorward expansion of either vortex ring should be accompanied by a parallel contraction or expansion of the other ring, but that the amount of the latitudinal shift of the anticyclonic vortex ring should be less than that of the cyclonic vortex ring. If the cyclonic and anticyclonic vortices are assumed to have equal and opposite rotation about their own axes, then the relationship is given very simply by

$$\cos^2\phi - \cos^2\phi' = K \tag{1}$$

where K is a constant of integration, and ϕ and ϕ' refer to the latitudes of the cyclonic and anticyclonic vortex rings, respectively.

The following statistical checks indicate that the above conclusions are not without some observational verification:

(1) In Fig. 6 are plotted curves showing the pressure difference in millibars between the max and min points on the weekly 5-day pressure profiles, and the corresponding latitudes ϕ' and ϕ of the max and min points of the profiles. On eight of these 43 weeks the pressure profile was not clear enough in outline to permit a certain choice of both the max and the min point which defined the west wind zone. These questionable points on the curves are marked by small circles.

(2) For the thirty-five weeks having pressure profiles with clearly defined max and min points, the correlation between ϕ and ϕ' is found to be $+0.66$.

(3) For the same thirty-five weeks, the correlation between ϕ' as observed, and ϕ computed from the formula (1) above by introducing the observed value of ϕ', also is found to be $+0.66$.

(4) Twenty-one of the above mentioned thirty-five weeks show a change in Δp from the preceding week of 3 *mb* or more, which was arbitrarily assumed to represent a significant change in the intensity of the zonal westerlies. For these twenty-one weeks the correlation between Δp and the current $\Delta\phi$ (change in latitude of the min point on the profile) was found to be $+0.55$, while the correlation of Δp with the current $\Delta\phi'$ (change in latitude of the max point on the profile) was found to be only -0.07. The significance of the first correlation, and insignificance of the second, is in agreement with the conclusions mentioned above which follow from the assumption of a tendency towards conservation of angular momentum in annular rings of the zonal circulation.

(5) The correlation of Δp with $\Delta \phi$ and $\Delta \phi'$, respectively, *one week later*, is found to be +0.34 and +0.46. These correlations show a significant tendency for an increase in the zonal westerlies to be followed later by a poleward shift of the entire west wind zone, and vice versa. A comparison of these correlations with (4) above shows definitely that the shift appears first in the low pressure trough at high latitudes, and is followed by the high pressure belt at low latitudes.

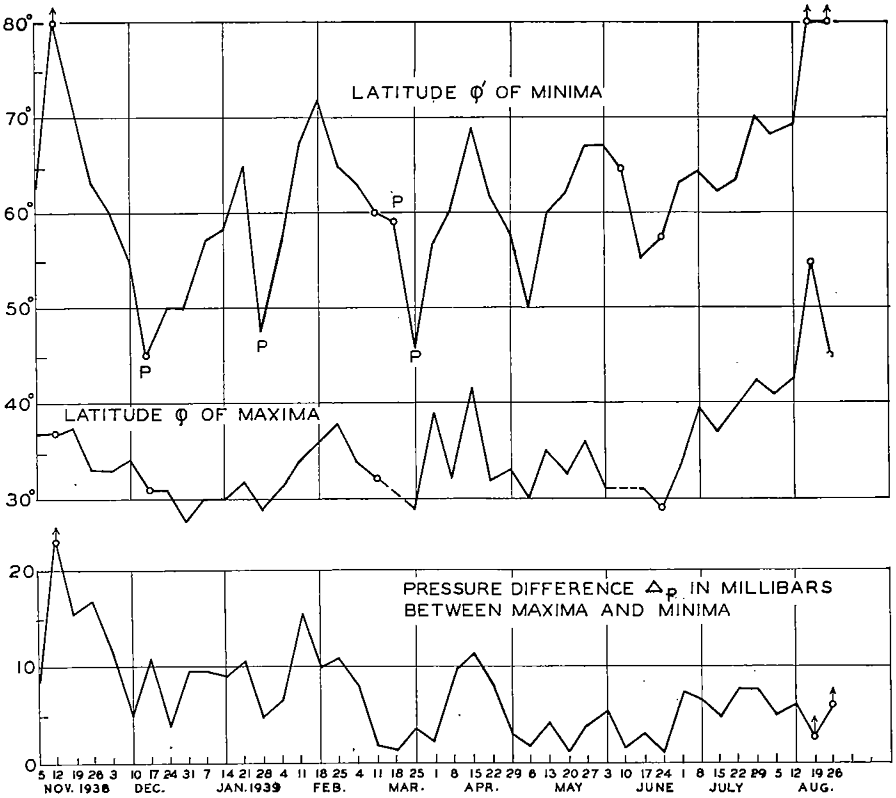

Fig. 6. Variation in the latitude of maximum and minimum points on the weekly mean Northern Hemisphere pressure profile.

6. *Relation of the north–south transport to the zonal index*

It is apparent from the velocity profile shown in Figure 4 (p. 26) that on either side of the belt of westerlies a region of considerable wind shear exists in the mean. It has been suggested by Rossby and Pekeris* that a strong horizontal wind shear of this sort probably is unstable, and by means of lateral mixing would result in the dissipation of the energy of the westerlies into large-scale horizontal eddies. That is, if the westerlies are strong one week, there should be a tendency for increased large-scale eddy motion

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* Transactions of the American Geophysical Union, 19th Annual Meeting, 1938.
"Fluid Mechanics Applied to the Study of Atmospheric Circulations," this series, Vol. VII, No. 1, 1938.

the following week. Increased eddy motion should be shown by increased meridional transport in the mean, which in turn should be represented by stronger east–west and west–east pressure gradients. An index of the average strength of such pressure gradients was computed by averaging around the hemisphere for each 5° latitude circle, the pressure gradient along the circle, representing the north and south components of the gradient wind. The average was obtained by summing without regard to sign the pressure differences along the latitude circle from maximum pressure to minimum pressure, minimum to maximum, etc. The sum of the pressure differences was then multiplied by a factor to convert to mean geostrophic wind velocity, with the appropriate correction for latitude and average temperature. These meridional indices were computed daily for the period November, 1938, to May, 1939, for every 5° latitude circle from 20° to 75°N, and averaged to obtain consecutive 5-day means for comparison with 5-day mean zonal indices. The correlation of the meridional indices for different latitudes with the zonal index does not indicate any significant relation, either contemporary or at lags up to two weeks.

A further test was made which would allow particular comparison with the meridional index the following week of those cases in which a steep velocity gradient already existed on the current velocity profile. On each zonal velocity profile was plotted the change in the mean meridional velocity at each 5° latitude circle from the given week to the following week. It was then possible to see in each case of strong westerlies whether a large zonal velocity gradient existed, and if so whether it was followed a week later by an increase in the mean meridional velocity at the same or any other latitude. No quantitative measure of such tendency was attempted since it was apparent from examination of the charts that no consistent relation existed.

7. *Correlations of the zonal index with its quadrant parts*

It is obvious that the latitude and intensity of the semi-permanent centers of action will have considerable to do with the zonal index. For instance, the mean pressure difference will be greatest when the Lows are deep and located on the 55° latitude circle, and when the Highs are strong and located on the 35° circle. In order to determine the relative effect of land and ocean areas on the index, the average 35°–55° pressure difference was obtained for four separate sections, the Atlantic, Asia, the Pacific and North America, comprising respectively 70°W to 10°E, 20°E to 130°E, 140°E to 140°W, and 130°W to 80°W. Each section was correlated with all other sections and with the hemisphere index for the 33 weeks from October, 1938, to May, 1939, with the results shown in the table below.

TABLE II

	ATLANTIC	ASIA	PACIFIC	NORTH AMERICA
Atlantic		.13	.19	.39
Asia			−.20	.06
Pacific				.25
Hemisphere	.77	.44	.53	.62
Hemisphere (expected)	.51	.45	.45	.28
Probability	.02	>.9	.55	.02

A positive coefficient would be expected for each correlation of a section with the hemisphere index because part of the variation of the hemisphere index is due to the particular section. By the "expected" correlation is meant the correlation that would be obtained if the sections were mutually independent. It is equal to the ratio of the standard deviation of the sectional index to the standard deviation of the hemisphere index. The correlations of the Asiatic and Pacific sections with the hemisphere index are little different from the expected values, whereas, due to their comparatively high mutual correlation (0.39), the Atlantic and North American sections are significantly correlated with the hemisphere index. The variation in the hemisphere zonal index thus is most closely associated with the North American and Atlantic sections, that is, an index based on only these two regions will be a good estimate of the index for the hemisphere as a whole.

This fact is significant because, since the outbreak of the war, the difficulty of obtaining data has made it necessary to eliminate the Asiatic quadrant and the eastern third of the Atlantic quadrant in computing the zonal index. Even this index, which is computed at the Weather Bureau in Washington, has not been available at M.I.T. to be used currently in the preparation of the five-day forecasts. For this purpose the zonal index from 60°W to 160°W has been used, covering only North America and the adjacent areas of the Atlantic and the Pacific. Such an index, however, is found to correlate rather closely with the Northern Hemisphere index, and probably can serve the same purpose in forecasting for the United States. The correlation between this North American Index and the Northern Hemisphere index, for two periods totaling 39 weeks during the colder half of the past two years was found to be 0.74. The correlation would probably be less during the summer, but the relative weakness both of the indices and their fluctuations in summer make them comparatively useless for forecasting.

There is also some indication that much of the time there exists a definite lag correlation between the Northern Hemisphere zonal index and the North American. At the beginning of the 1939–40 winter season this was very striking. The correlation for the first thirteen weeks of this season between the zonal index (exclusive of Eurasia) and the North American index one week later was found to be 0.85 which was produced principally by one extreme eight week period. The same correlation for 24 weeks of the colder half of the preceding year was found to be 0.42. A longer series of winter values of these two indices is not yet available for correlation, but probably the smaller of the two coefficients above indicates as much as could be expected from a long series. However, any basis for forecasting the North American index will be useful in making five-day temperature forecasts in the eastern U. S., as is indicated in a later section of this report. (See p. 56.)

There are several other relationships involving the zonal index and the local circulation patterns in the different longitudinal quadrants which have been investigated statistically. One such correlation tried was that between the zonal index and the intensity of the Bermuda High (as defined by the pressure at 35°N, 65°W). This latter quantity is important for temperature anomalies in the eastern U. S. (See p. 57.) A positive correlation was expected, because strong subtropical highs contribute to, and may in part be a result of, a strong zonal circulation. However, for 19 seasonal mean pressure maps for the fall and winter seasons from 1921–1930, the correlation was not particularly striking, only 0.33.

A second, more significant relationship involving the zonal index and the local quadrant circulation patterns, one which has received considerable attention, is that between weekly variations in the index and the behavior of some of the principal centers of action of the general circulation. From certain theoretical considerations concerning the general circulation which are outlined by Rossby,* an expression for the velocity of the semi-permanent perturbations in terms of their wavelength and the zonal wind velocity may be obtained. If C is the velocity of the perturbation (positive toward the east), then

$$C=\frac{U-L^2\beta/4\pi^2}{I+L^2/4\pi^2\lambda^2}$$

where U is the zonal wind velocity, L is the wavelength of the perturbation, β is the rate of change of Coriolis parameter toward the north, and λ is a constant, assuming a homogeneous, incompressible atmosphere. For a given stationary wavelength and given value of U, if the zonal velocity decreases, c becomes negative and the perturbation will move eastward. Once the center has moved from its normal position the presence of solenoidal fields held stationary by topographic factors may tend to generate a new center in the normal position, thus giving double centers when circulation decreases. Rossby* shows two examples verifying the above theory. In one example the correlation between the zonal index and the longitude of the center of the Aleutian Low as determined from 5-day mean maps is found to be nearly .70 for a period of about 5 months. The other example shows a movement to the westward and a splitting of the Asiatic High during a pronounced decrease of the zonal index.

The changes in longitude of the Aleutian Low show a pronounced correlation with changes in the zonal index for the past three winters. The correlations are:

WINTER	CORRELATION	MEAN INDEX
1936–37	0.60	5.9 mb
1937–38	0.28	3.5
1938–39	0.70	6.0

Further examination has been made of the relation of the index to the movements of other centers. An attempt was made first to obtain positions of the centers and simply correlate the longitude of the position with the zonal index. For the oceanic Highs especially this is unsatisfactory, since the flat gradient in the center of a High makes it impossible in most cases to specify with certainty the center, and a movement may appear on the map as a ridge of high pressure without a displacement of the central isobar. That may account in part for the small correlations found. Such a movement may be very significant in determining weather conditions at a given locality, however. A good example is the effect of a westward extension of the Asiatic High on the weather of Northern Europe, which will be discussed in more detail a little farther on.

Since the displacements of the centers should be most pronounced after the strongest changes in the zonal index, the Northern Hemisphere charts for maximum and minimum index should then show characteristic differences. For strong circulation the stationary wavelength of a perturbation is greater, hence it would be expected that in the Pacific and Atlantic there would exist single high pressure cells of large area, and that the Aleutian and Icelandic Lows would exist as single centers, probably to the east of their normal positions. During weak circulation, on the other hand, the Pacific and Atlantic Highs should be small or indistinct centers, and the Aleutian and Icelandic Lows move west of

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* "Relation between Variations in the Intensity of the Zonal Circulation of the Atmosphere and the Displacement of the Semi-permanent Centers of Action," *Journal of Marine Research,* Vol. II, No. 1, 1939.

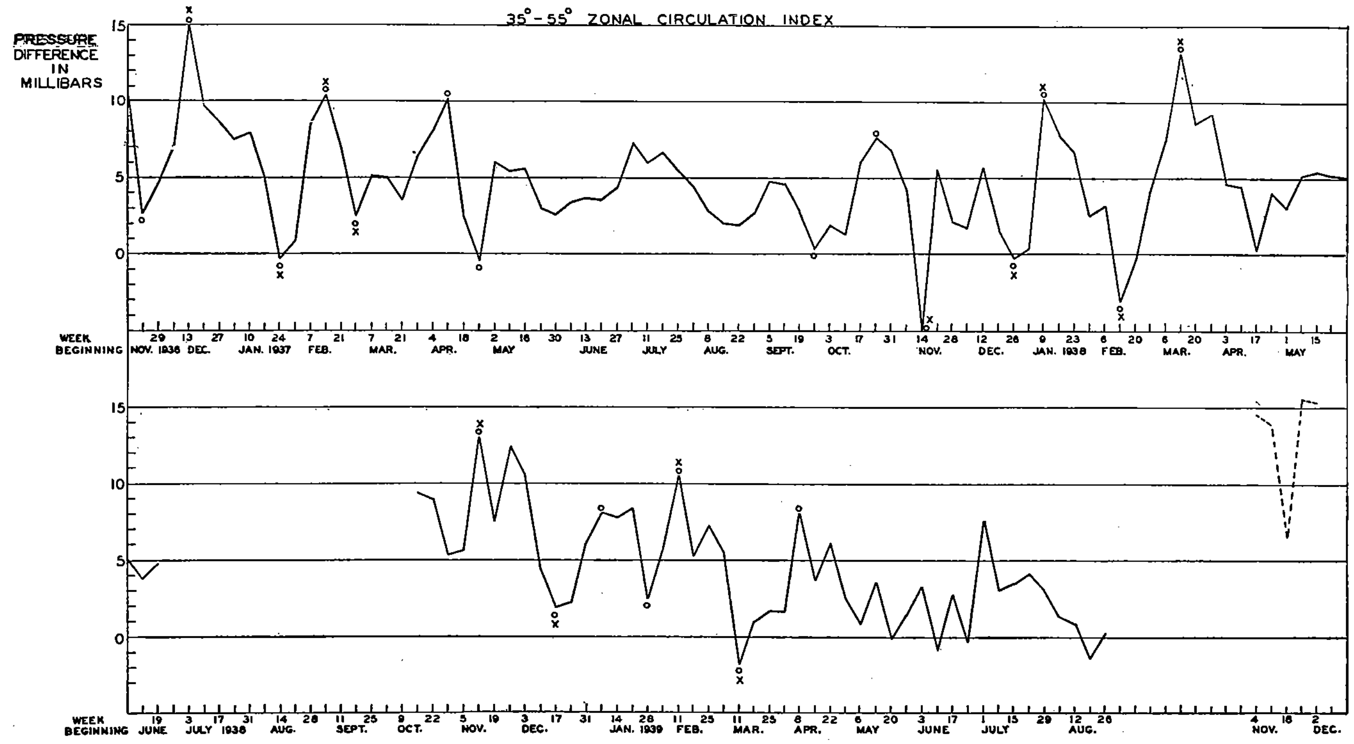
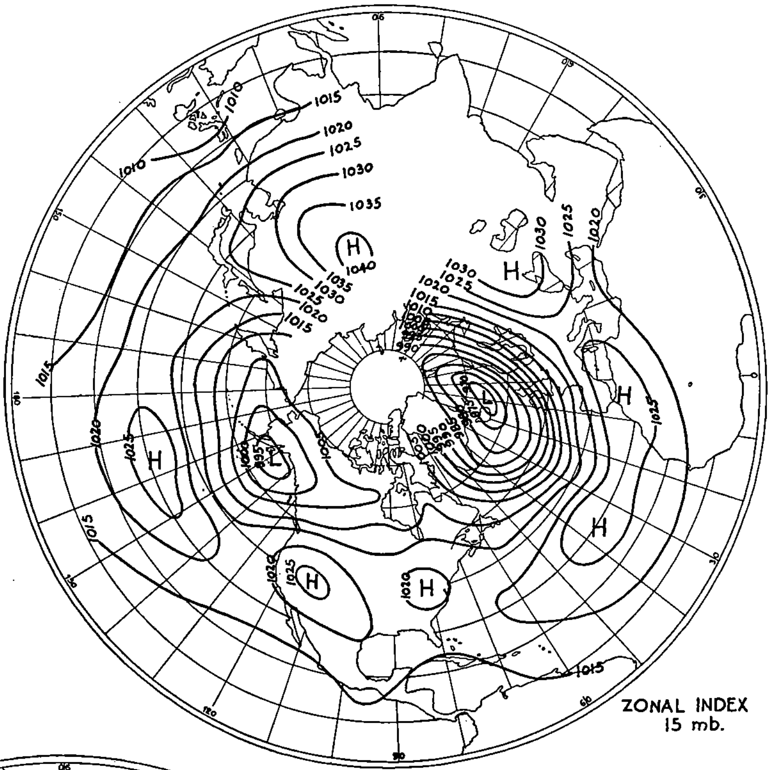


FIG. 7. Weekly mean pressure difference between 35° and 55°N latitude for November, 1936 to June, 1938. 5-day mean for October, 1938 to August, 1939.

their normal positions, with possibly a second center in the normal position. Figure 7 shows the series of zonal index values from which periods of maximum and minimum index were selected. The points marked by crosses are those for which charts are shown in the following figures. Note that each point of maximum index follows a large and prolonged rise in the circulation, and each minimum follows a large and prolonged fall. Only the periods during which winter conditions predominate were used, for the band of westerlies is much weaker in summer and fluctuates more irregularly. In general the winter period is taken as the months October through March.

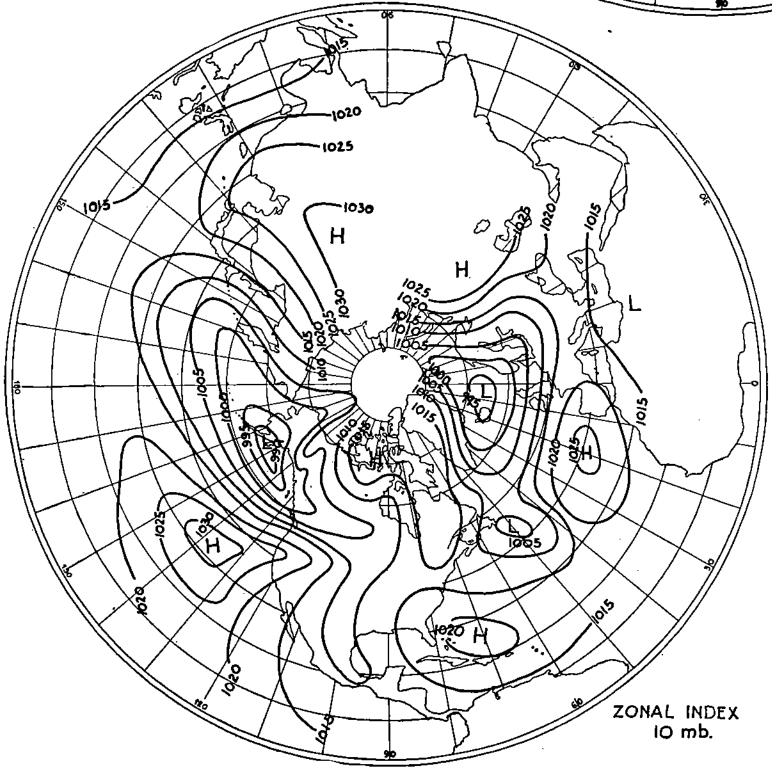
Figures 8–13 show Northern Hemisphere mean pressure maps for the periods of maximum index. The 35°–55° mean pressure difference in millibars is given at the lower left. In Fig. 8, showing the mean pressure for the week of December 13–19, 1936, after a strong rise in the index for two weeks, the Aleutian Low is a strong single center, and is somewhat east of its normal position. The Icelandic Low also is well-developed and in the eastern Atlantic. The Atlantic and Pacific Highs are both single large centers. From the spacing and size of the centers it is evident that the wavelength of the perturbations is quite long.

Figure 9 shows the mean pressure for another period of strong circulation, February 14–20, 1937. Note that the Aleutian Low and Pacific High are both well-developed single centers near the coast of North America. The Icelandic Low has a weak secondary center in the western Atlantic, but the stronger center is about in its normal position. Figure 10, showing mean pressure for January 9–15, 1938, has again a single deep Aleutian Low.



ZONAL INDEX
15 mb.

FIG. 8. Northern Hemisphere mean pressure, December 13–19, 1936.



ZONAL INDEX
10 mb.

FIG. 9. Northern Hemisphere mean pressure, February 14–20, 1937.

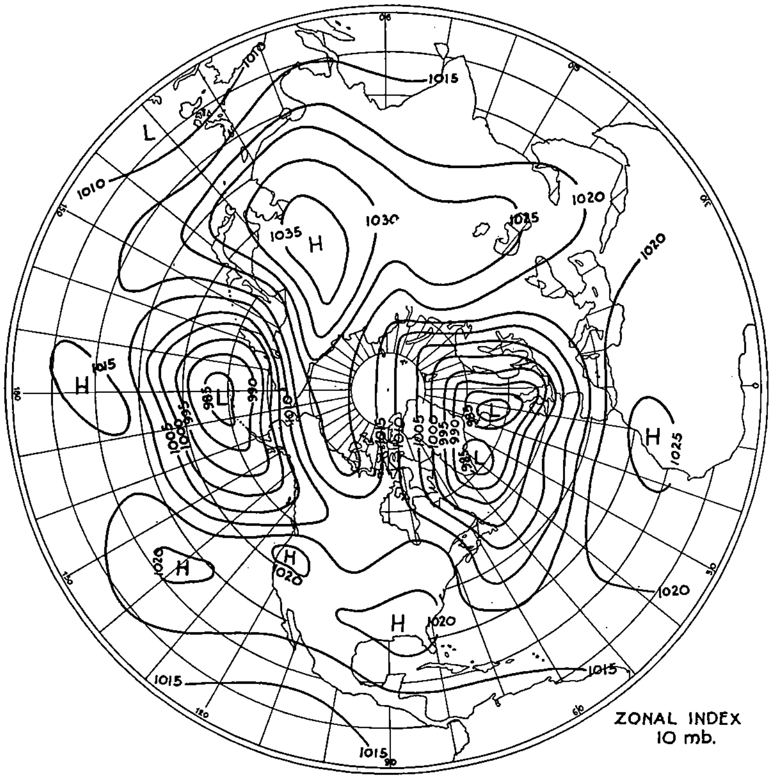


Fig. 10. Northern Hemisphere mean pressure, January 9–15, 1938.

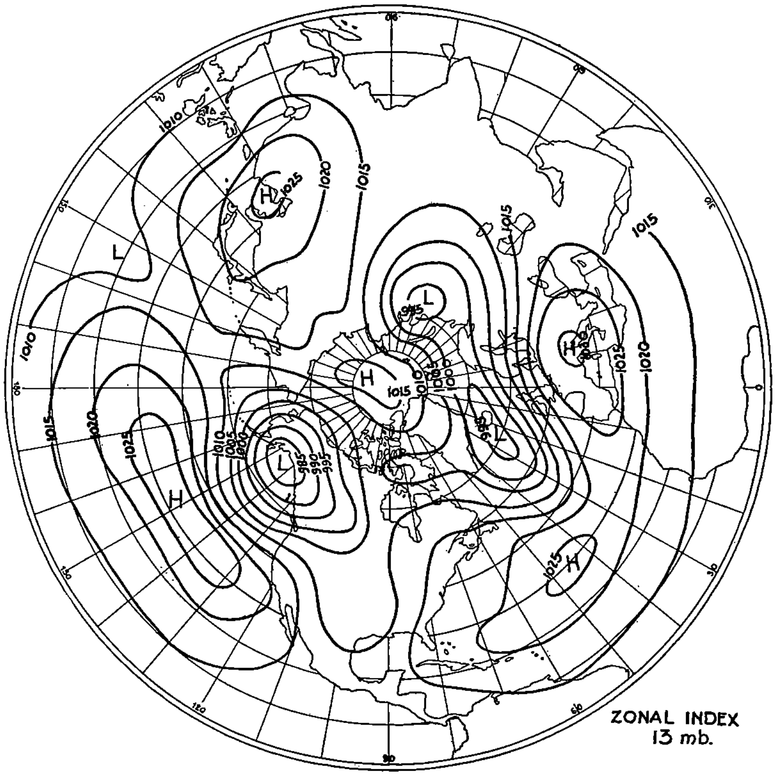


Fig. 11. Northern Hemispher mean pressure, March 13–19, 1938.

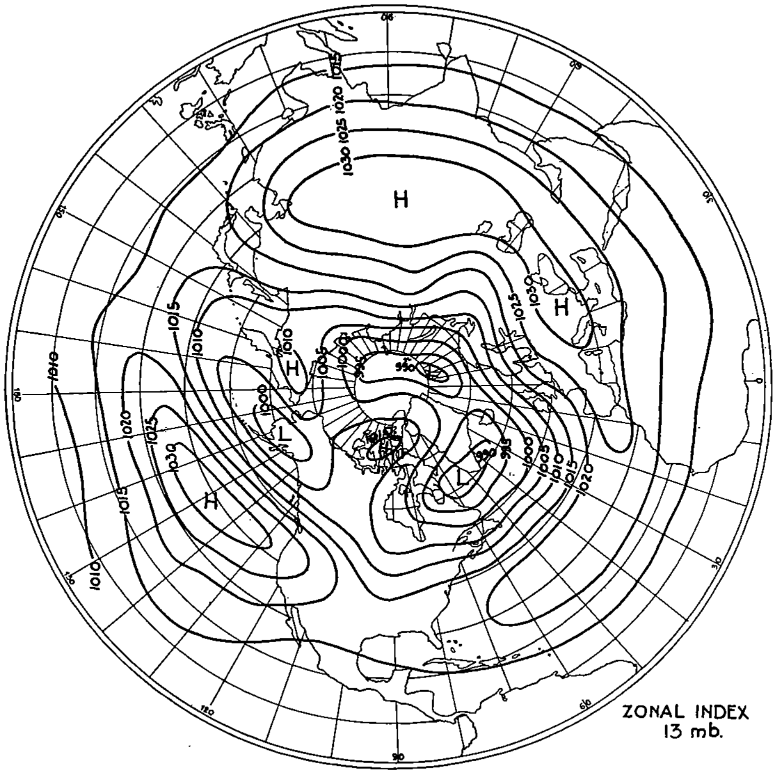


Fig. 12. Northern Hemisphere mean pressure, November 12–16, 1938.

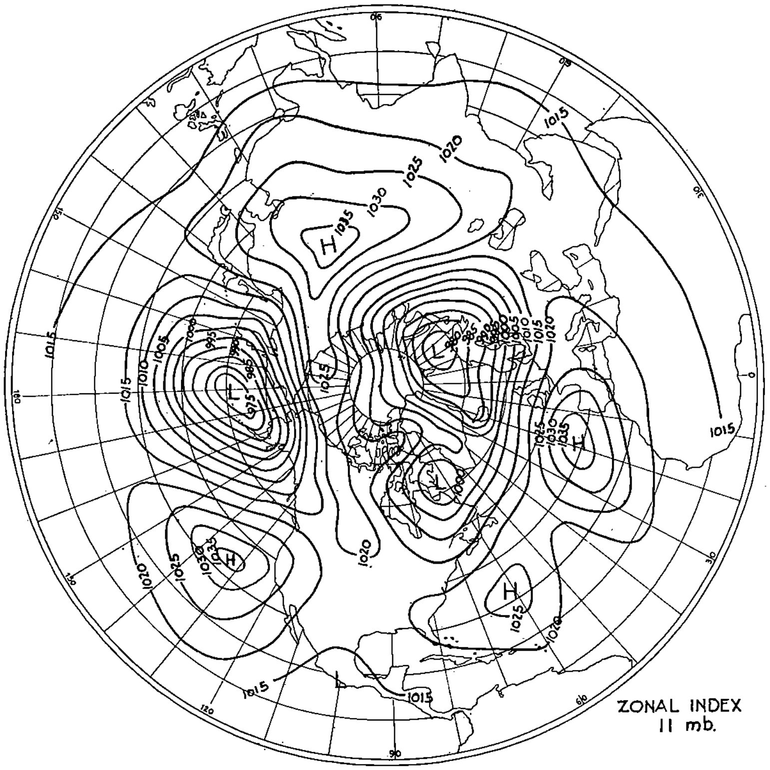


Fig. 13. Northern Hemisphere mean pressure, February 11–15, 1939.

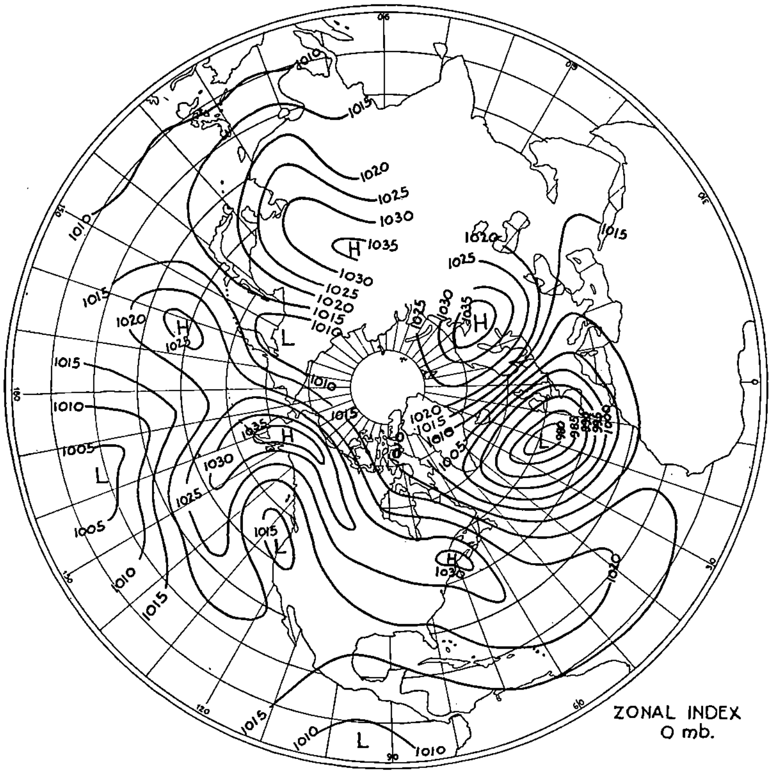


FIG. 14. Northern Hemisphere mean pressure, January 24–30, 1937.

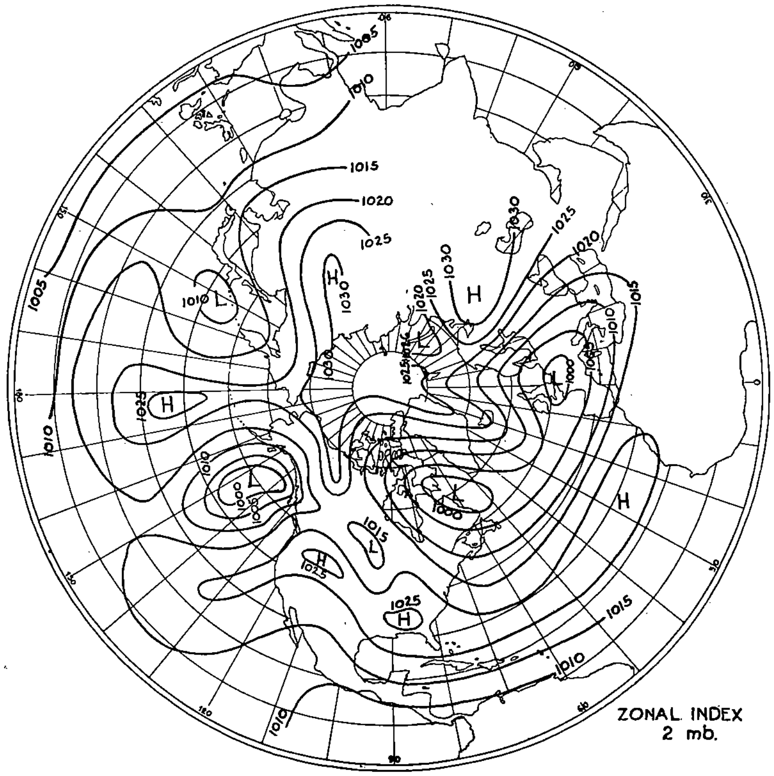


FIG. 15. Northern Hemisphere mean pressure, Feb. 28–Mar. 6, 1937.

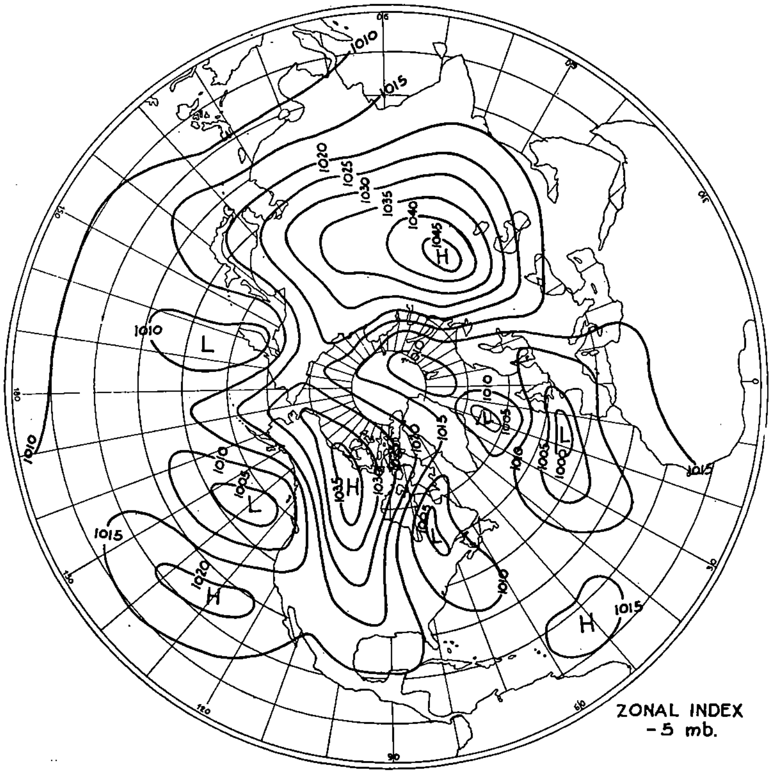


Fig. 16. Northern Hemisphere mean pressure, November 14–20, 1937.

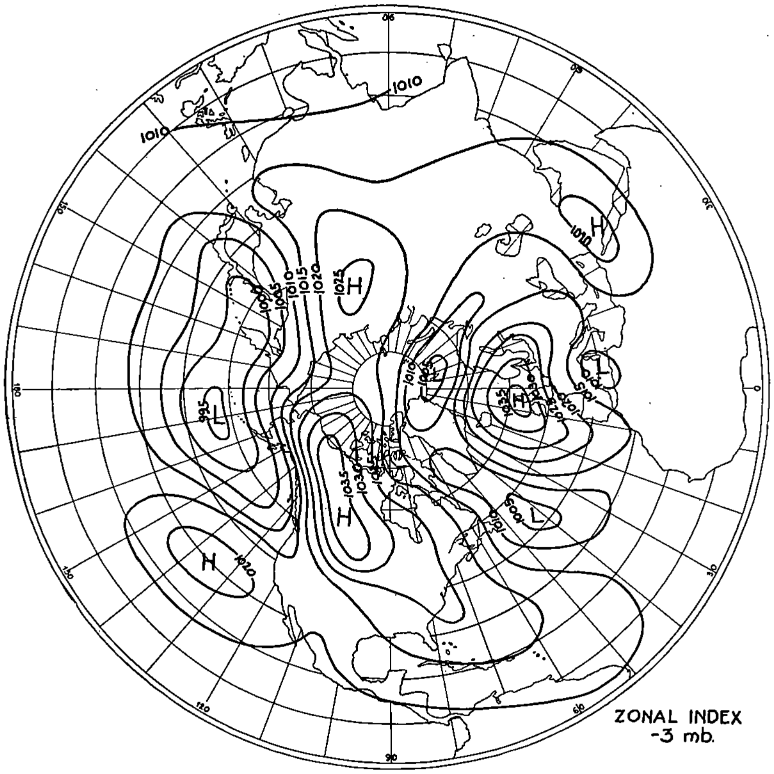


Fig. 17. Northern Hemisphere mean pressure, February 13–19, 1938.

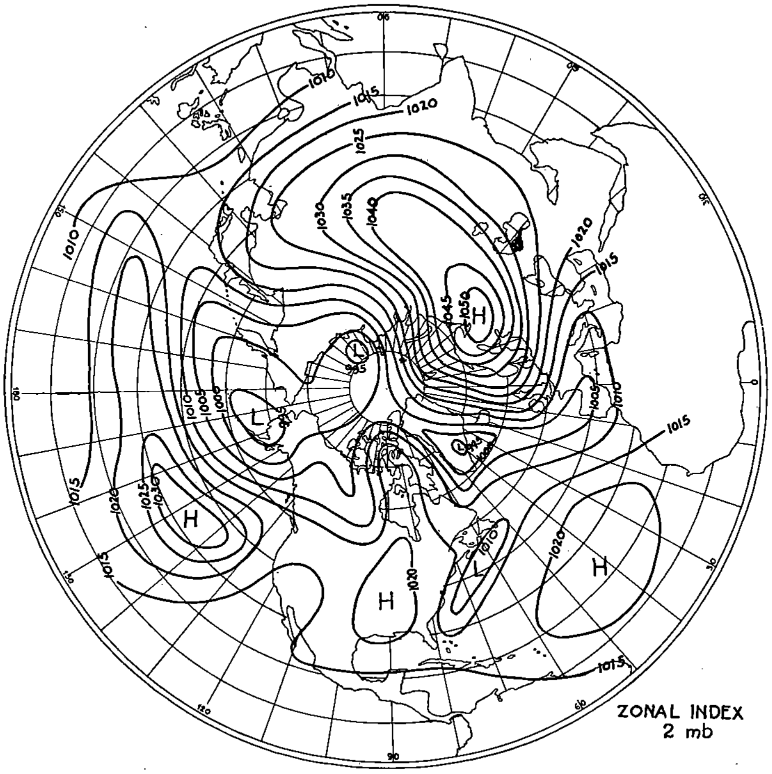


Fig. 18. Northern Hemisphere mean pressure, December 17–21, 1938.

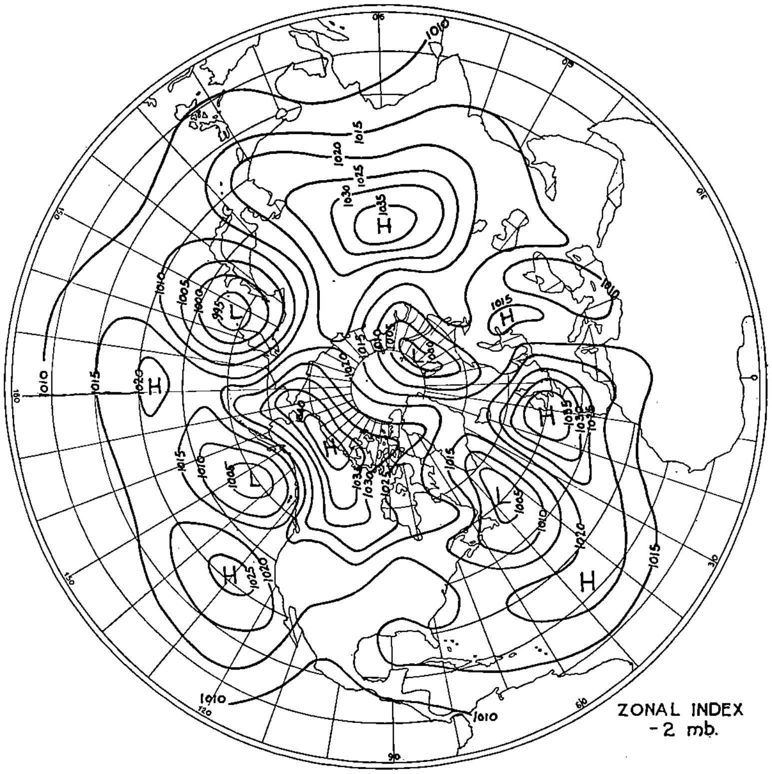


Fig. 19. Northern Hemisphere mean pressure, March 11–15, 1939.

The Icelandic Low is deep and about in its normal position. The apparent tendency to split into two centers was due only to a slight irregular shifting from day to day of the single deep Low. Figures 11, 12, and 13 give further evidence that during maximum index the Aleutian Low exists as a single center east of its normal position, and the Icelandic Low as a strong center dominating the circulation over northern Europe. In each case also, the Asiatic High is confined to eastern Asia.

In contrast to the circulation pattern for strong index is the pattern for weak circulation, as shown in Figs. 14–19. In Fig. 14, showing the mean pressure pattern for the week January 24–30, 1937, it is immediately apparent that the circulation is broken into much smaller cells than with strong index. The Aleutian Low is split into two weak cells, and the air flow over the west coast of North America is off-shore instead of on-shore, producing cold, dry weather. The Icelandic Low is far south of its usual position, and the Asiatic High has pushed westward, bringing a flow of continental air instead of maritime air over northern Europe. There are no true Atlantic or Pacific Highs.

During the week of February 28–March 6, 1937 (Fig. 15), after a decrease in the index, the circulation pattern contains many small centers instead of a few large centers. Here the Aleutian Low and Icelandic Low are both split into two centers, as is also the Asiatic High. Figures 16–19 show further examples of westward displacement and breaking up of the centers of action. In Figs. 16, 17, and 18 note the particularly strong Asiatic High, far to the west of its usual position.

On four of the six charts for minimum index (Figs. 14, 16, 17, and 19), there exist strong high pressure centers over Canada. On none of the charts for maximum index is there such a center. For North America this pronounced tendency for weak circulation to be accompanied by a building up of high pressure over Canada is one of the most important characteristic differences between maximum and minimum index.

Another of the differences which is important for the weather of the west coast of North America is the splitting of the Aleutian Low during minimum index. When the Low is a well-developed single center in its normal position there ordinarily is only one frontal system in the Pacific, with wave-cyclones occluding in mid-Pacific, and the occluded fronts moving inland along the northern coast. Precipitation in such a case is confined to the northern coast in general. However when the circulation index reaches a low value, with a split Aleutian Low, there will be found two paths along which cyclones are moving, one in the western Pacific and one in the eastern Pacific. The eastern Low center is usually farther south than normal, and the precipitation associated with the frontal system falls along the southern coast.

On the six mean pressure charts for maximum index, there are altogether 15 Low centers, or an average of 2.5 per chart. On the charts for minimum index on the other hand, there are altogether 25 Lows (omitting those above 75° and below 30° latitude), or an average of 4.2 per chart. For the Highs during maximum index the average number is 4 centers, and for minimum index the average is $4\frac{1}{2}$. Thus there is a definite measurable tendency for the circulation to be broken into more and smaller centers with minimum zonal index, that is, for the distance between centers (the wavelength of the semi-stationary perturbations) to be smaller during weak than during strong zonal circulation.

It may be noted from Fig. 7 (p. 36) that during each of the three winters examined there has been one period during which the index decreased steadily for several weeks, and simultaneously the Asiatic High moved westward, bringing continental air and low temperatures for a prolonged period to northern Europe. These periods were January

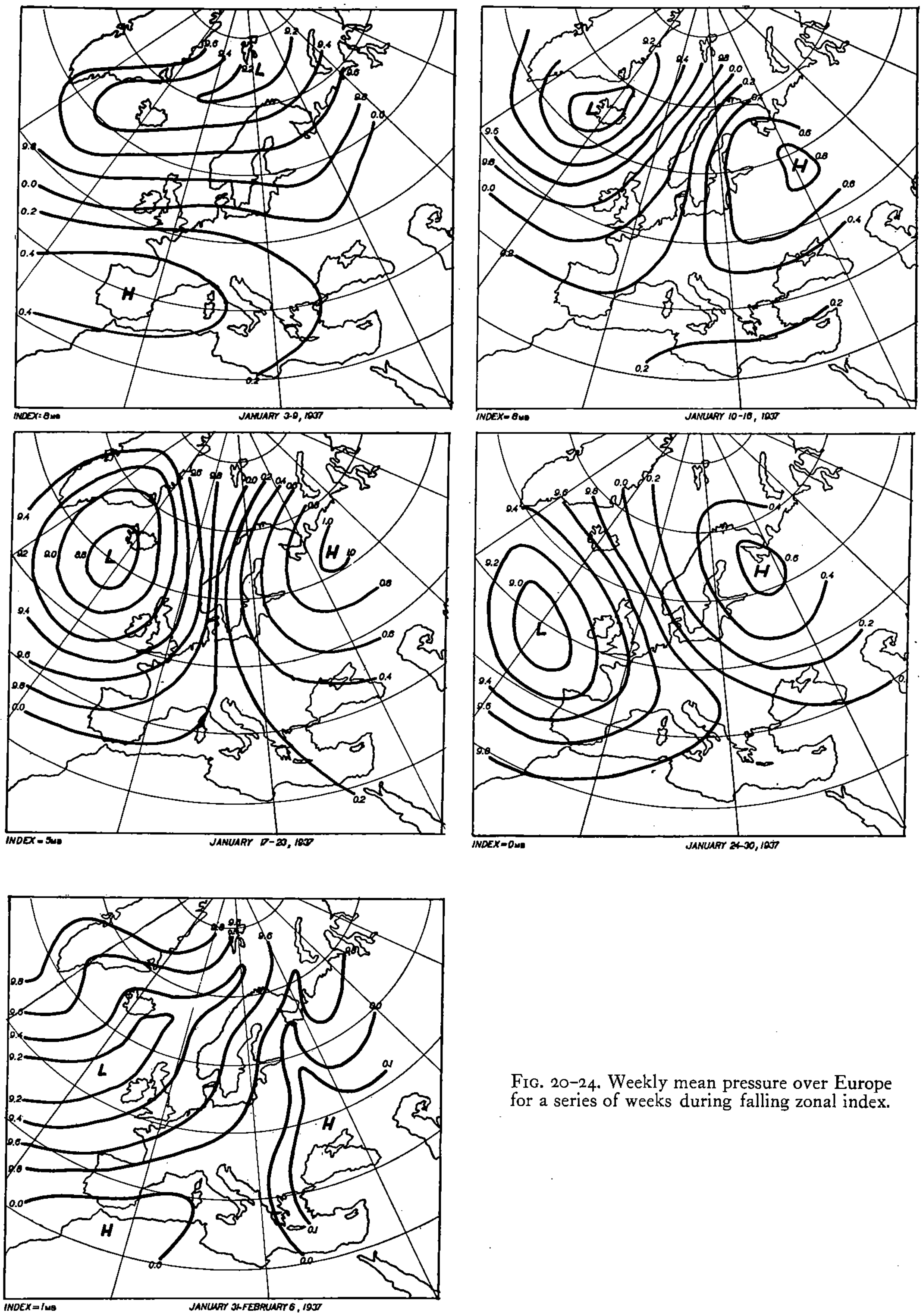


FIG. 20–24. Weekly mean pressure over Europe for a series of weeks during falling zonal index.

10–30, 1937; January 16–February 15, 1938; and December 3–24, 1938. Figures 20–24 show the sequence of mean maps for the period in January, 1937. Previous to the week of January 3–9 (Fig. 20), and during part of this week the Icelandic Low had been bringing warm, maritime air over Europe. On January 8, a High moved in, joined with the Asiatic High on January 10, and remained in about the same region for three weeks (Figs. 21–23), bringing exceptionally cold air from Siberia westward over all of northern Europe. It was not until February (Fig. 24) that an inflow of warm air from the Atlantic began to moderate the temperatures, in response to an eastward movement of the Icelandic Low and a recession of the Asiatic High to the eastward. Weekly mean charts for the periods in January, 1938 and December, 1938 show similar cold waves. For the period in December, 1938 the Asiatic High was particularly strong (Fig. 18), and record low temperatures were reached at many points in northern Europe. Ability to forecast the beginning and time of duration of these pronounced decreases in the zonal index would be of immense value in predicting severe cold waves in Europe.

Other relationships involving the zonal index and the local quadrant circulation patterns refer to the North American quadrant, and specifically to the conditions in the upper atmosphere as shown by the three kilometer mean pressure maps and the mean isentropic charts. The three kilometer chart in winter, and the isentropic chart in summer, are two of the most useful tools available for the preparation of five-day forecasts in the United States.

In connection with the three kilometer pressure map, the first characteristic of practical forecasting significance is the intensity of the latitudinal pressure gradient at that level. It is a good indicator of the general trend of the west-east movement of the day to day weather changes currently and in the near future. It was pointed out in a preceding section that the 35°–55° zonal index is as good a measure of the strength of the westerlies as can be obtained from surface pressures. It might be expected that a similar index computed from upper level pressures (e.g., a three kilometer chart) would be a better index, for it would be free from the influence of surface disturbances. For the index to represent the mean westerlies, it would have to be computed for the whole hemisphere. Nevertheless, some correlations were computed between the surface zonal index and a similar index obtained from the North American three kilometer chart, to determine whether any tendency toward a relation exists. The lack of observational data restricts the three kilometer index to the pressure difference between 35° and 45° latitude, averaged for the longitudes 70° to 120° W. Using five-day means for the period October, 1938 to April, 1939, no correlation with the current zonal index was found. The three kilometer chart from which this index was taken represents so small a section of the westerlies, however, that lack of correlation should be taken merely as an indication that this three kilometer index is not representative of the average upper level circulation. This fact renders more difficult a forecast of this index but does not lessen its indicative significance once it is known.

But even more important than the three kilometer index in its forecasting significance is the longitudinal position and movement of the trough of low pressure which is normally present in winter on these weekly mean three kilometer pressure maps. (Figs. 30, 31, p. 61.) On the east side of this trough the general trend of movement of the day to day weather changes is northward, with above normal temperature, on the west side southward, with below normal temperature. Since the presence of this trough in the three kilometer isobars corresponds to an important perturbation in the westerlies, it was hoped that its longitudinal displacement from week to week might be correlated with the zonal index, in accordance with Professor Rossby's criterion for a stationary perturbation. (See p. 35.) Such correlation would furnish a basis for forecasting the displacement of this important phenomenon. However, the correlation of the apparent eastward movement of the three kilometer mean pressure trough with the zonal index for the 38 weeks in which this trough was reasonably well defined during the colder half of the past two years, yielded a coefficient of only -0.13. This coefficient is both insignificant and of the opposite sign to that expected. The probable reason for the failure of the expected correlation between zonal index and the apparent perturbation movement is indicated in the discussion of the similar correlation with the North American three kilometer index. (See p. 59.)

Finally, there were investigated possible relationships between the zonal index and the quasi-stationary anticyclonic eddies of the summer mean monthly isentropic charts. The intensity and position of these quasi-stationary upper level eddy circulations is very significant for the summer weather conditions in the United States. (See p. 70.) Hence it is most important to consider their driving mechanism and any factor which may help in forecasting their characteristics. According to Rossby,* the westerlies of the middle latitudes furnish the chief supply of energy for the anticyclonic cells which are found along their southern border. In summer there are few solenoids over the southern United States, and this state of quasi-barotropy extends roughly to the Canadian border where the normal Polar Front is located. Here is concentrated the field of solenoids which is associated with the prevailing westerlies. In winter, on the other hand, the polar front is farther south, and a study of resultant winds over the United States at this time of year brings to light the well known fact that the southern limit of the westerlies, and hence the anticyclonic cells, are displaced southward beyond the range of our aerological network. The summer season, therefore, permits a detailed study of the structure and development of the frictionally driven quasi-stationary anticyclonic eddies.

According to Rossby† there should be a relation between the size of these anticyclonic cells, or quasi-stationary perturbations, the zonal velocity of the westerlies, and the latitude. Making certain assumptions Rossby computed the wavelengths of the perturbations which should be stationary for various latitudes and zonal velocities. He called these quantities "stationary wave lengths," and here these may be assumed to be comparable to the diameters of the quasi-stationary anticyclonic cells. The average diameter of the large anticyclonic eddy observed over the southwestern United States in summer is about 3600 km. This value was obtained from the mean isentropic chart for the period containing the months June, July, August, 1934 to 1939. (See p. 71.) According to Rossby's theoretically computed values this calls for a zonal velocity of eight meters per second. Now the observed zonal velocities of the westerlies along the northern United States border (on the isentropic surface $\theta = 315°$ for which the normal isentropic chart was constructed) average about force 5 Beaufort, that is, from 8 to 11 meters per second. It appears, therefore, that Rossby's results are in fair accord with the observations.

In view of this agreement and the theory it might be supposed that the variations of

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* "On the Role of Isentropic Mixing in the General Circulation of the Atmosphere," *Proc. of the Fifth International Congress of Applied Mechanics*, 1938.

† "On the Mutual Adjustment of Pressure and Velocity Distributions in Certain Simple Current Systems," *Journal of Marine Research*, Vol. I, No. 1, 1937.

cell size which are observed from one monthly mean isentropic chart to another are related to the zonal index as measured from the Northern Hemisphere monthly mean surface pressure charts. Unfortunately, however, the dimensions of the anticyclonic cells cannot be rigidly established by the configuration of moisture and contour lines* and the resultant winds on the mean isentropic charts. An attempt was made to measure the diameter of the western cell by finding the distance between the most probable eddy center (determined by inspection of the monthly mean isentropic charts) and the axis of the dry tongue on the eastern periphery of the eddy. These values were then plotted for each of the summer months from 1934 to 1939, as were the zonal indices (35°–55°), but there appeared to be no correlation between them. Perhaps part of the reason for this result lies in the lack of a more definite measure of the cell sizes, but probably most of the difficulty lies in the fact that there are important factors other than the zonal index which should affect the cell size, notably the latitudinal rate of shear which exists within the westerlies, and the variable thickness of the air layer included between two isentropic surfaces. When the isentropic surfaces converge toward the north, which is frequently the case, anticyclonic vorticity is produced on the west side of the cell, which tends to reduce the cell size. Probably a quantitative investigation of these different control factors may help to explain the cell patterns, but a more extended aerological network is really needed for this purpose.

B. Correlation of the Zonal Index with Local Anomalies of Temperature

For the most part local anomalies of the meteorological elements are best correlated with the respective local quadrant indices and circulation patterns, rather than with the state of the general circulation as a whole. However, a few relationships of the latter type, notably with the zonal index of the general circulation, have been investigated. It might be added that there has been noted recently a promising indication of such a relationship between the winter precipitation on the west coast and the zonal index, which it has not been possible to check statistically in time for this report. There follows a brief discussion of a few such temperature relationships which have been investigated.

1. *Patterns of the temperature departure from normal with maximum and minimum values of the zonal index during the colder half year*

When the zonal circulation intensity falls to a low value in winter, it is generally accompanied, as shown in a preceding section (p. 43) by the building up of a high pressure center over Canada. Such formation of a high over the cold, snow-covered continent provides the necessary stagnation to produce a deep, cold P_c air mass, which then can be expected to drain southward over the United States, producing a major, prolonged cold wave. Thus large or prolonged decreases in the zonal index should be associated with periods of generally sub-normal temperature in the United States.

On the other hand, when the circulation is strong, the Aleutian Low is generally in the eastern Pacific, and maritime air is brought into the continent. There is no chance for prolonged stagnation, hence exceptionally cold air masses are not found. Probably the Gulf also contributes warm air to help maintain high temperatures in the eastern and

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* It now appears that charts of the isentropic stream function would aid considerably in this matter.

and southern states, but it is not intended to suggest here the circulation pattern by which this is accomplished. Thus above-normal temperatures should be associated with maximum values of the zonal index.

Charts of temperature departure for the United States have been examined for 17 weeks of minimum index and for 18 weeks of maximum index. Two charts for maximum index (December 16–22, 1936, and January 12–18, 1938) and two for minimum index (January 27–February 2, 1937, and March 15–21, 1939) are shown in Figs. 25–28. Figures 25 and 26 correspond to Figs. 8 and 10 (pp. 37, 38), respectively, which show the mean pressure for the two periods of maximum index. The temperature departures were obtained from the Weekly Weather and Crop Bulletin, and are weekly means beginning three days later than the mean pressure charts. Note that for these two periods of strong circulation the temperature departures were predominantly positive for the United States, and that the largest departures were in the northern and western states, where Pacific air penetrated the country.

Figures 27 and 28 show temperature departures for minimum index, corresponding to pressure charts in Figs. 14 and 19 (p. 40, 42), respectively. The low temperatures in Fig. 27 are confined to the western states, whereas in Fig. 28 the area of negative departure is east of the Mississippi River. Apparently a knowledge of other factors is necessary in order to determine whether, with decreasing circulation, the cold wave will occur in the east or in the west.

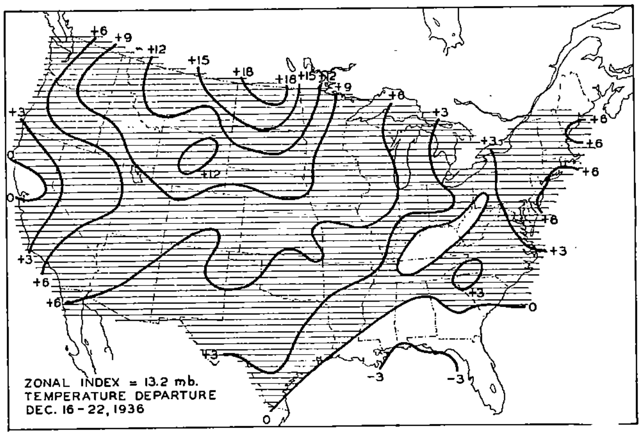

Fig. 25. Temperature departure for a period of maximum index.

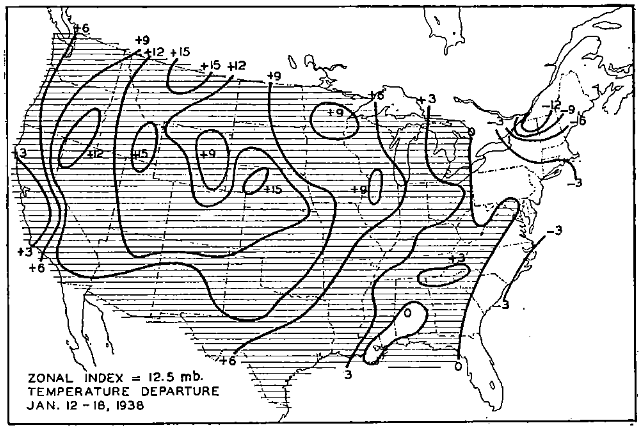

Fig. 26. Temperature departure for a period of maximum index.

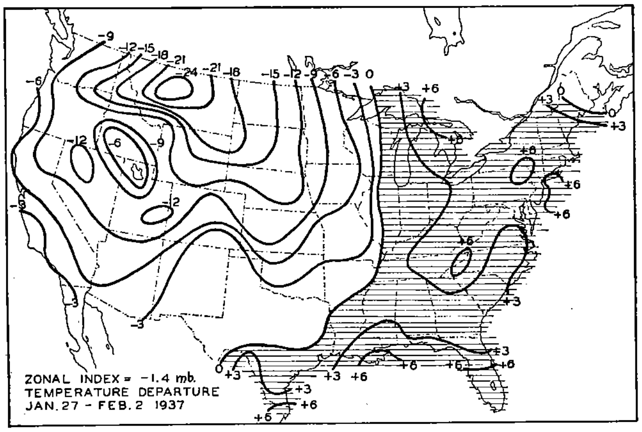

Fig. 27. Temperature departure for a period of minimum index.

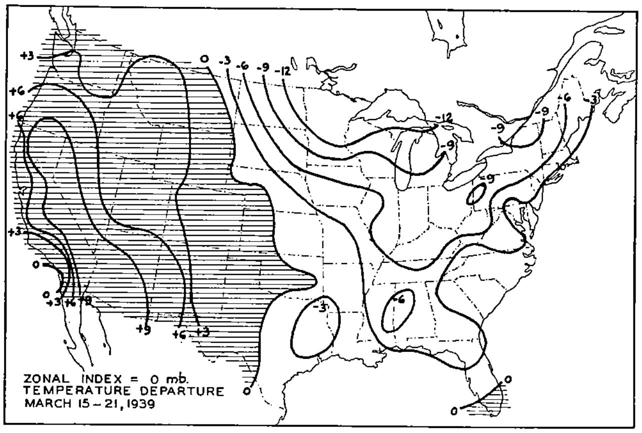

FIG. 28. Temperature departure for a period of minimum index.

Several of the periods of low circulation which were examined, particularly one in January, 1933, and one in December, 1938, were accompanied by predominantly high instead of low temperatures. Examination of the mean pressure charts for these two periods shows that the Canadian High failed to build up, hence there was no opportunity for the formation of a deep P_c air mass. All but one of the 18 periods of maximum index, on the other hand, were associated with temperature departures predominantly positive, particularly in the eastern United States. Thus low index is accompanied by generally low temperatures only when cold air is available from Canada, whereas high index is accompanied by high temperatures in the eastern states almost invariably, and in the western states also most of the time. The western states were above normal in eight of the nine cases when the index was above 10 millibars.

2. *The zonal index and temperature departures from normal in the eastern United States*

From the discussion of the preceding paragraphs it is easy to justify a correlation between the zonal index and temperature departures from normal in the eastern United States. This was checked first by correlating the weekly values of the zonal index with the current weekly departure from normal of the temperature in the eastern United States, as indicated by three practice forecast districts comprising New York state, northern Georgia and Alabama, and western Missouri and eastern Kansas. The correlation was positive, as would be expected, but insignificantly small for the past two winter seasons and that of the Polar Year. There seems definitely a tendency for major cold waves to be associated with small and decreasing values of the zonal index, but the point of cold air outbreak may be located in widely separated regions around the earth, and its occurrence may be more or less delayed relative to the occurrence of the minimum of the zonal index.

A further statistical check was made by correlating the change in the zonal index from one week to the next with the change of the temperature departure from normal in the eastern United States (as defined above) during the following week, i.e., allowing one week's lag for the temperature change. Such a lag is justified in view of the fact that the verification districts lie well on the south side of the circulation index zone, while the cold waves associated with a decrease of the zonal index must affect northern areas first, southern later. The correlation obtained was again positive, but only slightly more significant than that obtained above. It amounted to 0.44 for the first fourteen weeks of the past winter season, to 0.33 for 23 weeks of the preceding winter season, and to only 0.15 for 25 weeks of the colder half of the Polar Year.

It was considered advisable to correlate also seasonal values of the zonal index, and the corresponding seasonal temperature anomalies in the eastern United States, for the ten-year period 1921–30, for which the Weather Bureau had prepared seasonal mean pressure maps over the Northern Hemisphere. The temperature anomalies in the eastern United States were determined by the same three districts as in the case of the weekly values above. No lag correlations were computed here, for a lag of a whole season would not make physical sense in this relationship. Moreover, it should be emphasized that there is less reason to expect significant correlation between seasonal temperature anomalies and the zonal index than in the case of weekly values. Such a difference may be expected from the fact that whereas large-scale polar outbreaks have periods such that their southward progress individually may frequently control the weekly fluctuations of the zonal index, such control is scarcely possible in the case of seasonal variations of the index. Seasonal variations should be little affected by the southward progress of major polar outbreaks, for the reduction of the index normally associated with the inception of the outbreak should be compensated later by an increased zonal index when the outbreak has reached low latitudes.

In these seasonal correlations, the spring and summer seasons are always correlated independently from the fall and winter, by reason of the much smaller circulation activity during the first two seasons. In this case, the spring and summer correlation of the zonal index with the current seasonal temperature anomaly in the eastern United States was found to be +0.49 (twenty seasons). The corresponding fall and winter index (nineteen seasons) was −0.49. No explanation of these apparently significant coefficients of opposite sign is offered. It seems probable that they would break down for a longer series. Also correlated with the seasonal temperature anomaly in the eastern United States was the change in the departure from normal of the zonal index since the preceding season. (The normal of the zonal index for each season was simply taken as the arithmetical mean of the ten values for this ten-year period.) This coefficient was found to be +0.24 for the spring and summer seasons, and −0.44 for the fall and winter seasons. These change correlations are in line with those above of the index itself, but again it seems doubtful if much significance can be attributed to them.

C. Correlation of Sunspots with Local Temperature Anomalies

In view of the fact that there is some evidence of correlation between sun spots and the general circulation (p. 29, et seq.), it seemed advisable in correlating local temperature anomalies with other elements to include a few sunspot correlations. Coefficients were computed between the monthly sunspot numbers (Wolfer) and the monthly mean temperature departures from normal at Paris (maritime, western Europe), Albany, N. Y. (continental, eastern North America) and Portland, Oregon (maritime, western North America), for the four months December through March of the eleven-year period 1908 through 1918, giving thus 44 pairs of values at each station. The only significant correlation appeared at Paris, with a coefficient of −0.47. This negative correlation might be explained on the basis of a southward shift of the Polar Front (i.e., southward shift of storm tracks), during sunspot maxima. This would be in accordance with indications mentioned previously (p. 29). Unfortunately, however, when the same coefficient was computed for a fifty-year period (1873–1923) using first annual mean temperature departures and sunspot numbers, then again monthly values for the same four winter months, no significant correlation was found in either case.

A similar correlation was made between the seasonal sunspot numbers and the seasonal temperature departures from normal in the eastern United States (as defined on p. 49) for the ten-year period 1921–30. A negative correlation was found for the fall and winter seasons, but it was not large enough to be significant, considering the shortness of the period covered.

D. Interrelationships between Branches of the General Circulation

Up to this point only relationships involving the state of the general circulation of the northern hemisphere as a whole have been discussed. However, similar relationships involving only quadrant indices or individual centers of action may prove more significant, statistically, in some cases. There follows a discussion of a number of such local interrelationships between parts of the general circulation.

1. *Mutual correlations between the quadrant sections of the zonal index*

It was pointed out by Garriott* in 1904 that a slowing down of the eastward moving lows and highs often occurs first over the British Isles, to be followed about a week later by a slowing down over northeast Canada. In terms of the zonal circulation this should be represented by a lag correlation between the circulation index for a section representing the British Isles and the index for eastern Canada a week later. Garriott also suggested that speeding up of the motion over Canada was followed a week later by speeding up over the British Isles. If this is so, then a lag correlation between the index for the Isles and the index for Canada a week later will be diminished from what it would be if only the first effect were present. However, if the trend in the index tends to persist for periods considerably greater than a week, as has been shown to be the case, then a lag correlation should still show the presence of the first effect, and a correlation at a week's lag in the other direction should show the second effect.

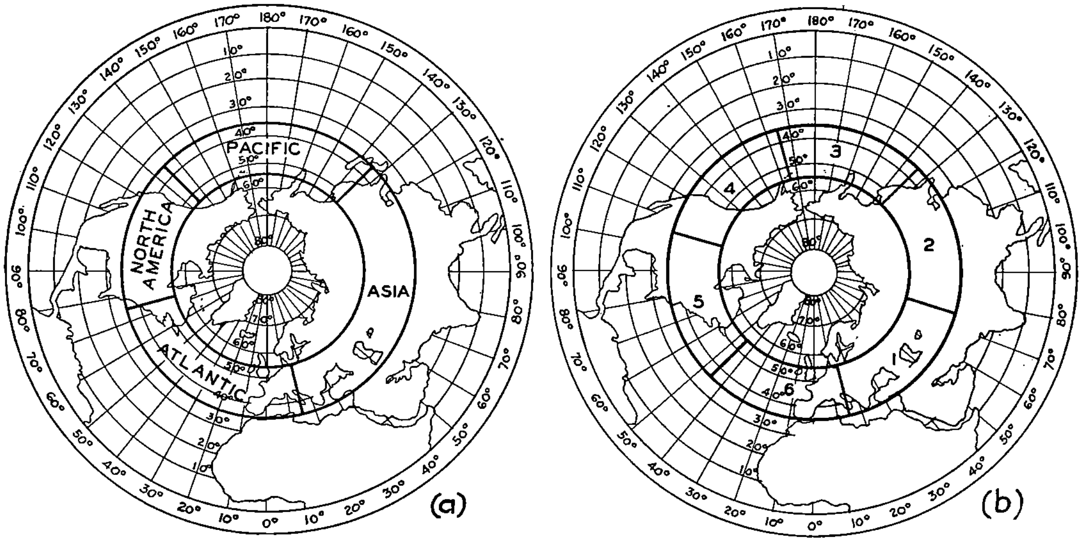

FIG. 29. Division of the hemisphere for sectional indices.

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* *Weather Bureau Bulletin*, No. 35, 1904.

The first correlations tried were lag correlations between the weekly indices for sections divided according to ocean and continent (Fig. 29a). That is, the index for Asia was correlated with the index for the Atlantic a week later, and with the index for North America a week later. Also the Atlantic was correlated with North America a week later, and North America was correlated with the Atlantic a week later. The correlations obtained are shown in the table below, together with current correlations between the sections. The 35°–55° pressure difference was used for all sections as a zonal index, although for the Atlantic and Asia an index for higher latitudes would represent the westerlies better. The periods included in the given years were:

1932–33	September 27, 1932	to	April 3, 1933	7-day means
1936–37	November 15, 1936	to	April 3, 1937	7-day means
1937–38	September 26, 1937	to	April 2, 1938	7-day means
1938–39	October 15, 1938	to	May 3, 1939	5-day means, weekly

TABLE III

	No Lag		One Week Lag		Two Weeks Lag	
Sections	1932–33	38–39	1932–33	38–39	1932–33	38–39
Asia-Asia				.54		
Asia-Pacific	.49	−.20		.02		
Asia-America	.03	.06	.28	.24	.42	
Asia-Atlantic	.24	.13	.49	.32		.10
Pacific-Asia				−.27		−.14
Pacific-Pacific				−.03		
Pacific-America	.44	.25		−.26		
Pacific-Atlantic	.23	.19	.27	.17	.48	.19
America-Asia				−.07		
America-Pacific			−.36	.10		−.16
America-America				.07		
America-Atlantic	.32	.39	.02	.24		
Atlantic-Asia				−.04		
Atlantic-Pacific			.20	.50	−.02	.15
Atlantic-America			.38	.18		.04
Atlantic-Atlantic				.39		

Correlation Between Index for First Section and Index for Second Section at Given Lag Later.

The correlation between Asia and the Atlantic a week later was 0.49 for the 1932–33 period, and 0.32 for 1938–39. Neither coefficient is very large, but both years suggest a positive correlation. The correlation of Asia with North America a week later is even smaller for both years, but is still positive. The correlation of Atlantic with North America a week later was comparatively large (0.38) in 1932–33, but smaller (0.18) in 1938–39. On the other hand, the correlation of North America with Atlantic a week later is small for both years, thus suggesting that the "blocking" travels westward rather than eastward. Other correlations were obtained as shown in the table. Many of the coefficients are fairly large, and for the two tables combined there are more large positive coefficients than would be expected in the same total number drawn from an uncorrelated population. However, the criterion as to the significance of a correlation should be not only the magnitude of the correlation but also whether it persists from year to year.

For further correlations the index was divided into six sections as shown and numbered in Fig. 29b. It was felt that the sections used above were too large, and that variations within the sections were as significant as variations between sections. The correlations between these smaller sections are shown in the table above. The one outstanding correlation is between section 6 and section 5 a week later. The correlation ranges from 0.50 to 0.20 for the four years. This indicates just the sort of blocking mentioned by Garriott —i.e., when the circulation decreases in the eastern Atlantic, it tends to decrease a week later in the western Atlantic and eastern North America. The correlation for all four years combined is 0.32, based on 92 weeks, and is highly significant, although it means that only a small part of the variation in 5 is accounted for by 6 the preceding week. The correlation of 5 with 6 a week later is too small to suggest any lag relationship between the velocity over Canada and the velocity over eastern Atlantic a week later.

TABLE IV

	One Week Lag				Two Weeks Lag			
Sections	1932–33	36–37	37–38	38–39	1932–33	36–37	37–38	38–39
1-4				.15				−.22
1-5				.16				.04
1-6	.27		−.30	.36				.19
2-1				.00				−.20
2-6			−.18	.39				.03
3-1				−.27				−.20
3-4		.14	.12	−.25		−.33	−.06	.64
3-5		.19	.05	−.22	.26	.39	.01	.39
3-6				−.02				−.03
4-1				−.06				−.08
4-3	−.45	.36	.34	−.05				−.22
4-5				.05				−.11
4-6				.10	.21			.24
5-1				−.16				−.15
5-3	−.17	.17	.22	.19	−.19	−.10	−.09	−.14
5-4	.45	−.18	.29	.00				.14
5-6		−.03	.11	.24				.37
6-1				−.13				−.29
6-3	.06	−.47	.31	.40	.05	−.38	−.18	.10
6-4	.05	−.14	.04	.23	.14	−.51	−.21	−.04
6-5	.50	.32	.20	.42				−.16

Correlation Between Index for First Section and Index for Second Section at Given Lag Later.

Some of the other combinations show high correlation for one season, but are not verified by trying other years, except for section 3 with 5 two weeks later. Here a correlation of zero for 1937–38 detracts from its significance; nevertheless, it is suggested that a change in the index over the western Pacific tends to precede by two weeks a corresponding change over eastern North America.

2. *Winter correlation between the Aleutian and Icelandic centers of action*

Since the Aleutian low and Icelandic low are the two principal centers of cyclonic activity on the Northern Hemisphere during the colder half of the year (they become weak and poorly defined during the late spring and summer), it seemed advisable to correlate their intensities during the colder season, as a check on the degree to which cyclogenetic tendencies in the general circulation may be common to the entire Northern

Hemisphere, or more locally determined within the area of one ocean. The fact that the quadrant index correlations listed above indicate little current correlation between the Atlantic and Pacific quadrants suggests at once that the correlation between the intensities of these two cyclonic centers is probably small. The correlation of the weekly mean intensities of these two centers for 28 weeks of the colder half of the Polar Year was found to be 0.24. From October to March (19 weeks) of 1938–39 the correlation was about the same, of little significance. Inspection of the data did not indicate any significant lag correlation. Hence it does not appear that there is any common control factor of significance in the development of these two primary centers of action.

The fall and winter seasonal means of the intensities of these same two cyclonic centers were correlated separately for the period 1921–30. These correlations were found to be +0.76 for the fall, and −0.60 for the winter. In view of the short period (only ten seasons each) and the difficulty of explaining any such seasonal reversal of the correlation, little weight is attached to these two coefficients. However, it may be mentioned that the factors which control seasonal anomalies of these two centers may be quite other than those which control the weekly fluctuations, so that a more extended check on seasonal means would seem to be in order. It is important that the behavior of these two centers is of great significance for the determination of the fall and winter weather in western Europe and western North America.

3. *Intensity of the Aleutian Low and the Plateau High*

The development of the so-called Plateau High in the western United States is a practically important feature of the fall and winter weather west of the Rockies which has long been associated in the minds of some forecasters with the activity of the Aleutian Low. The closeness of the current correlation has been checked for both weekly and seasonal means (inspection of the data suggests very little lag correlation of any kind). The intensity of the Aleutian Low (defined as the lowest pressure in the Pacific east of 180°) was correlated with the intensity of the Plateau High (defined as the pressure at 40°N 115°W) as given by the five-day or weekly mean pressure maps. For forty-four weeks during the colder half of the past two years (1938–39 and 1939–40) the correlation was −0.50. For 28 weeks during the Polar Year (1932–33) it was only −0.11. However, there seems to be a significant relationship here. The same correlation was made for the fall and winter seasons (total 19) of the 1921–30 period. In spite of rather small seasonal fluctuations of the Plateau High, this coefficient was found to be −0.73. From this it appears that the correlation is significantly greater for the seasonal than the weekly means, although the series is too short to base final conclusions on these results. It will be seen below (p. 56) that the daily fluctuations of these two centers are even less closely correlated than the weekly means.

4. *Correlations between daily anticyclonic activity over North America and cyclonic activity over the adjacent oceans*

The discussion of the preceding paragraph suggests the possibility of correlation between cyclonic activity on the one hand, and parallel anticyclonic activity on the other hand. If such relationship is indicated in weekly or seasonal means, then there is reason to look for it also in the day to day pressure variations. Lag correlations of real significance for daily forecasting should be considered as a real possibility, even though such lag is not sufficient to be evident in the longer period means. Any such relationship which might be established for the daily fluctuations would bear indirectly also on longer range forecasting, in so far as it might throw light on the closeness and nature of atmospheric interactions. American forecasters have remarked in the past not only on apparent relationship between cyclonic activity in the Aleutian region and anticyclonic development in the western United States, but also on similar possibilities between cyclonic activity in the North Atlantic and anticyclonic activity in the eastern United States. Since no rigid statistical check of these possibilities appears to have been made up to the present, it seemed advisable to include something along that line in this investigation.

Considering the Pacific situation first, the simplest procedure is to select on the basis of past records a definite geographical area as the seat of the Aleutian Low, another area as the seat of the Plateau High, and then to correlate the pressure changes in these two areas as indicative of the behavior of the two centers of action. Correlations of this type are likely to be small, because actually from day to day the positions of the two centers of action vary widely from the normal location. On the other hand, the forecasting value of such correlations, if they are significant, would have a maximum of value because the correlations refer to fixed geographical areas.

The average of the pressures at Dutch Harbor and Kodiak, Alaska, were chosen for the Aleutian Low index and a similar average at Salt Lake City, Utah, and Winnemucca, Nevada, for the Plateau High index. For sixty consecutive days of November and December, 1938, the correlation between these two indices was −0.16, which is of the expected sign (intense low with a well-developed high), but insignificantly small. The correlation of the 24-hour change of the two indices for the same sixty days was only −0.10 (high intensifies as low deepens), which is entirely insignificant. The same correlation taking the Plateau High change one day later (some lag might be expected) gave a coefficient nearly zero. It was thought that perhaps a closer correlation would be obtained if it were restricted to days of strong intensity of the Aleutian Low in the designated region. From November 1936 through February 1939 there were 45 days on which the Dutch Harbor-Kodiak pressure was below 985 millibars. Using these days for the same correlations, there appeared to be almost no correlation between the pressures taken either currently or with the High at one day lag. The changes showed a correlation of +0.26 taken currently, and −0.14 when the High pressure change was taken one day later. This indicates a possible lag between the deepening of the low and subsequent anticyclogenesis in the Plateau region, but the correlations are too small to be significant. It may be concluded that at least for this selection of points there is no significant point with point correlation between the daily fluctuations of pressure in the Aleutian Low region and the Plateau High region.

A second possible procedure consists in correlating the fluctuations of the minimum pressure of the Aleutian Low with those of the maximum pressure of the most closely associated continental high. The Aleutian Low was restricted to the Pacific between 160°E and 140°W, and the Plateau High to the North American continent between 140°W and 100°W. This procedure should give higher correlation, if any exists, because the full cyclogenetic and anticyclogenetic effect is included and the advective effect is eliminated, but it would be of less forecasting value, relatively, because of the uncertainty in its geographic application to a specific forecast region.

This correlation of the Aleutian Low with the Plateau High was evaluated for the 56 days of the November-December, 1938, period for which both centers were well defined within the specified longitudinal zones. The correlation of the intensity of the low with the contemporary 24-hour change of intensity of the high was −0.20, and that of the change of the low with the change of the high was −0.25. This indicates a small tendency, not significant, for a deep and intensifying Aleutian Low to be accompanied by an increasing Plateau High. The same correlations were computed for those 28 of the 56 days on which the central pressure in the Aleutian Low was below 975 mb. In these cases of more intense cyclonic activity the above two coefficients become −0.54 and −0.56, which are quite significant

It may be concluded from the above correlations that there is definitely a relationship between the daily fluctuations of the Aleutian Low and the Plateau High such that an intense and deepening low tends to be accompanied or shortly followed by an increasing Plateau High. The point to point correlation is not significant, however, nor is the center with center correlation significant enough to have forecasting value unless restricted to the cases of more than average intensity of the low. In general this daily correlation appears to be less pronounced than the weekly correlation, which in turn appears to be less than the seasonal.

On the eastern side of North America the principal center of cyclonic activity in the western Atlantic (between 80°W and 40°W) was correlated with the principal center of anticyclonic activity on the eastern side of the continent (100°W to 60°W). It was desired to check the assumption that cyclogenesis off the Atlantic coast is associated with anticyclogenesis (building up of high pressure in the intensified Polar outflow) on the eastern side of the continent behind the deepening low. Again the period of November and December, 1938 was investigated, during which period there were 49 days with two centers well enough defined for correlation. There was found to be almost zero correlation between the intensity of the low and the contemporary 24 hour change in the intensity of the high. The correlation between the 24 hour change in intensity of the low with the contemporary change of the high was found to be +0.20, which is insignificantly small, and indicates negative correlation between cyclogenesis and anticyclogenesis. Inspection of the data gave no indication that either by use of lag or by selection of cases only of marked cyclonic activity would any significant correlation be found in this data. Furthermore, the stability of these pressure centers is not such that they can be identified on the mean pressure maps and correlated for longer periods like the Aleutian Low and Plateau High. Hence it may be concluded that there is no such correlation between cyclogenesis off the east coast and anticyclogenesis inland as appears to exist on the west coast. This may perhaps also be taken to indicate that in general in middle latitudes strong cyclonic removal of atmosphere in one region is compensated by anticyclonic accumulation of atmosphere in advance (to the east) of the deepening cyclone, and not in the rear.

E. Correlations between Branches of the General Circulation and Local Temperature Anomalies and between Local Temperature Anomalies

1. *The North American quadrant index and the temperature departure from normal in the eastern United States*

The North American quadrant of the zonal index might be expected to correlate more closely with the temperature departure from normal in the eastern United States than does the zonal index itself. The temperature departure was determined in the same manner as previously (see p. 49) from the three practice forecast districts. Correlation of the weekly values of the American quadrant index with the current temperature departure from normal during the colder half of the year produced the following coefficients:

1932–33	+0.44	(24 weeks)
1938–39	−0.004	(25 weeks)
1939–40	+0.66	(17 weeks)

The one year 1938–39 is considerably out of line with the other two. No explanation can be offered for this discrepancy. If the American index is correlated with temperature departure one week later, the following indices are obtained:

1932–33	+0.53
1938–39	+0.14
1939–40	+0.63

The year 1938–39 still is out of line with the other two, but certainly these coefficients are significant. Especially the lag relationship is useful in five-day forecasting.

Unfortunately no seasonal correlations of these quantities could be made, for the American quadrant index has not been computed for the 1921–30 period. It may be worth noting about these temperature departures from normal in the eastern United States that neither the weekly nor seasonal departures show any appreciable serial correlation, so that persistence tendency or change tendency is of no assistance in forecasting this element.

2. *Intensity of the Bermuda High and temperature departures from normal in the northeastern United States*

During the past two winters one of the principal considerations in making five-day temperature forecasts for the northeastern United States (New York State practice forecast area) has been the expected behavior (position and intensity) of the Atlantic subtropical high pressure cell. It has been considered as axiomatic that a well-developed Atlantic cell centered well over on the western side of the Atlantic should mean warm weather along the Atlantic coast, while a trough of low pressure over the western Atlantic should favor subnormal temperature on the Atlantic coast. To check this empirical rule statistically the pressure at the point 35°N 65°W was correlated with the current temperature departure from normal in the New York State area. Again it is recognized that such a point with point correlation will not give a full measure of the relationship, for the pressure at one selected point is by no means a complete index of the position and intensity of the Bermuda High. But if the relationship is at all close, this correlation should indicate the fact.

For 25 weeks during the 1938–39 cold season and 17 weeks of the 1939–40 season, the correlation of the weekly mean values of these two quantities was found to be 0.48, which, in view of the incompleteness of the pressure index, was considered to be a very satisfactory verification of the empirical rule. However, when the same correlation was made for the Polar year (27 weeks) the coefficient obtained was −0.12. No explanation is offered for the failure of the relationship during that year. It runs directly counter to all the recent synoptic experience, both in daily and longer range forecasting.

The same correlation was computed between the seasonal means of these quantities, for the fall and winter seasons of the 1921–30 period. The coefficient was found to be −0.22. This is not particularly surprising, however, as an entirely different relationship from that holding for the weekly fluctuations may be expected. That can be seen from the following simple consideration. Major polar outbreaks in the eastern United States usually accompany low pressure off the coast, but are normally followed by anticyclogenesis off the coast and a contemporary reaction to a high temperature on shore. When such fluctuations occur with a period of about two weeks for one cycle, a high positive correlation will be obtained above. However, when averaged over a whole season, it may well be that a season with numerous cold waves in the east (subnormal temperature) may result in a normal or slightly above normal Bermuda High, and consequently lead to a small negative seasonal correlation of the two quantities. In the same way a season subnormal in cold waves in the eastern United States might be characterized by a slightly subnormal Bermuda High.

3. *Correlation of winter temperatures in the far northwest (Montana) with those in the northeast (New England)*

During the last decade there have been several instances when the monthly temperature departures from normal in the northeastern United States and in the far northwest have been large and of the opposite sign. This has led to the assumption, in some cases, that there is a persistent tendency for such a contrast to occur. To check this, the monthly temperature departures in Montana were correlated with those in New England, for all of the winter months (December through March) on which either departure exceeded 6°F, from 1915–1937, inclusive. Forty pairs of values were thus obtained, and the correlation was found to be nearly zero. Inspection of the monthly mean maps indicated that whereas there are many winter months when persistent polar outbreaks on one side of the continent are associated with equally persistent tropical currents on the other side (to give the temperature anomaly contrast noted above) there are about equally many periods when the Polar outbreaks take a more central route, so that the country is cold both east and west, or when the Polar front and cyclonic activity are located further to the north than usual, so that most of the country is dominated by air masses of subtropical origin with above normal temperature. Consequently there is little or no correlation between New England and Montana temperature correlations in the long run. Temperature anomaly contrasts between those regions depend entirely on the general circulation pattern prevailing over the continent and adjacent oceans at the time.

4. *Correlation of winter monthly temperatures at a few selected stations*

It seemed advisable to make a rough check as to the extent to which the large scale and persistent polar outbreaks and tropical currents in North America and Asia may tend to occur in unison or in opposition on the two continents. A correlation of sorts was made by comparing the monthly departures from normal of the mean temperatures at Albany, N. Y. (eastern North America, continental), Portland, Ore. (western North America, maritime), Ochiai, Japan (eastern Asia, 47°N 143°E, continental) and Paris, France (western Europe, maritime). The correlations were computed for the 44 cold months (December–March) of the years 1908–1918, inclusive. None of the coefficients were very large, although that of -0.42 between Albany and Ochiai is probably significant, indicating that the eastern continental type of Polar outbreak or Tropical flow tend to occur during different periods, not simultaneously, on the two continents. Negative coefficients of -0.30 and -0.26 were found between Albany and Portland and Portland and Paris, suggesting a slight tendency to opposition of phase between those regions. Positive coefficients of +0.27 and +0.18 appeared between Portland and Ochiai and Albany and Paris, respectively. From this it appears there is a slight tendency for opposite sides of an ocean to be in phase in their monthly temperature anomalies, and for opposite sides of a continent to be out of phase. The strongest tendency is for the two continents to be out of phase. However, the few coefficients computed here, although they are all in line with this generalization, represent extremely slim evidence on which to base such a statement.

F. Relationships between Upper Air Conditions over North America and Either the Branches of the General Circulation or the Anomalies of the Meteorological Elements in the U. S.

The five-day forecast practice, especially in the absence of a complete northern hemisphere map, has come to depend principally on the indications of the upper air charts over North America. Much of this practice, like weather forecast practice in general, is based on a combination of experience and feeling for the synoptic situation as a whole which cannot be specifically formulated and statistically checked. In the following discussions both statistical and synoptic evidence has been used to justify or to discredit certain hypotheses.

1. *Correlation of the North American zonal index with the 3-km index*

The significance of the zonal and quadrant indices of the state of the general circulation has been discussed at several points, but it was found that a similar zonal index at the level of 3 kilometers in the U. S. showed little correlation with the northern hemisphere index. (See p. 45.) Since it is important to determine the degree to which the mean surface pressure gradient indicates the mean motion of the atmosphere above, the correlation was computed between the North American surface index and the 3-km index between 35°–45° over North America. Lack of more aerological data has made it impossible up to the present to extend this check to a larger territory.

To make this check, the 35°–45° pressure difference was computed from the surface map for the same longitudes as the 3-kilometer index, namely 70°W to 120°W. The correlation with the 3-kilometer index for the 33 weeks in the winter of 1938–39 was 0.35, which is considerably higher than the correlation with the zonal index for the whole hemisphere. However, the smaller the longitude band used in the comparison, the smaller the correlation is likely to be. This correlation indicated that when the surface zonal index is broken down into sections smaller than 60° of longitude, it begins to lose its ability to represent the general flow aloft. It is also probable that over the oceans there would be, especially in winter, a much closer correlation between the surface index and the 3-kilometer index in the same region, than over continents. This should follow from smaller local frictional and thermal disturbances of the lower atmosphere by the surface beneath.

2. *The 3-kilometer index and the motion of the 3-kilometer pressure trough*

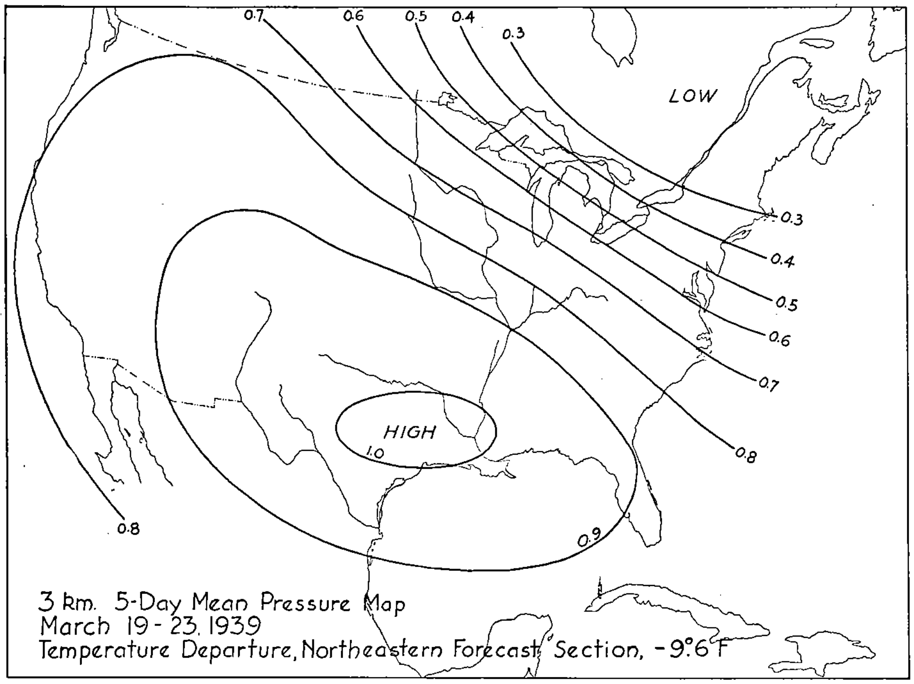
The significance of the position and movement of the trough of low pressure normally present in winter on the 3-kilometer pressure map over the northern U. S. as a general control of the weather regime in the eastern part of the country has already been mentioned. (See p. 45.) The correlation of the apparent eastward movement of this trough with the zonal index produced an insignificantly small negative coefficient instead of the positive correlation, which was anticipated. If the same correlation is computed using the 3-kilometer index instead of the zonal index, then there can be no doubt but what the coefficient obtained refers specifically to the intensity of the zonal flow in the immediate vicinity of the perturbation. For this correlation coefficients of −0.28 and −0.36 were obtained for the cold seasons of 1938–39 (24 weeks) and 1939–40 (14 weeks). Although these coefficients are not significantly large, still there can be little doubt that the correlation is negative.

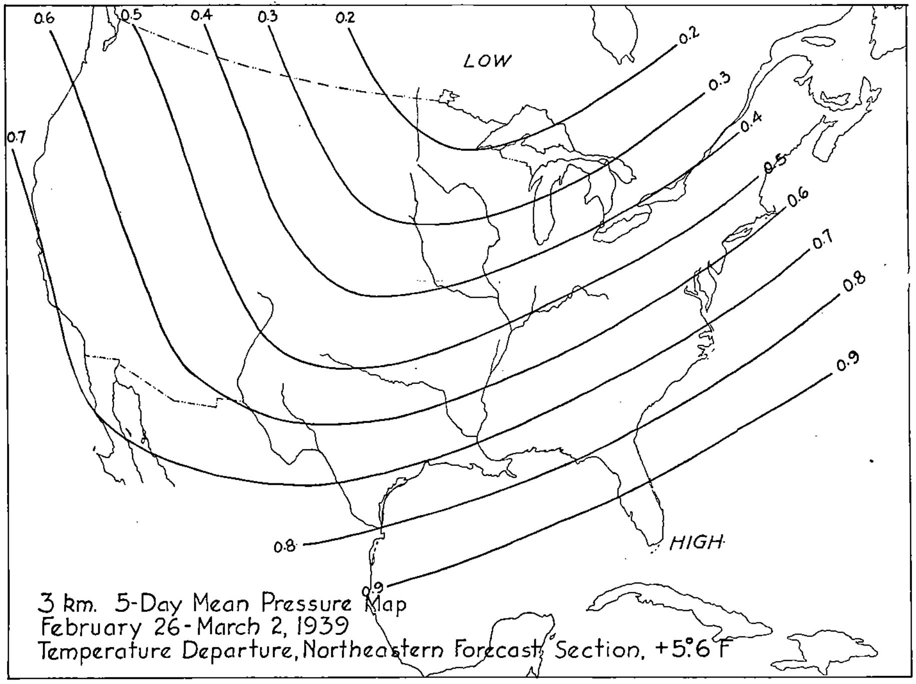
Some indication of the reasons for this unexpected result is suggested by an inspection of the weekly mean 3-kilometer maps. In the first place, since there is no overlapping of the 3-kilometer mean pressure maps, it is impossible definitely to identify these statistical pressure formations from one week to the next. The correlation computed here, and the similar one for the zonal index (p. 46), refer to the position of the effective 3-kilometer pressure trough in the United States, but it is not possible to say, in many cases, whether the position of the same trough is being compared, or whether perhaps a new trough which has arrived from the Pacific has established itself in the same region. In fact, the Northern Hemisphere mean pressure maps give indications, in some cases, that this is just what does occur when the zonal index decreases and the upper level pressure trough becomes established (apparently moves) further west. The preparation of overlapping mean maps, especially if they could be extended further out over the oceans, should clear up this question.

In the second place, an inspection of the 3-kilometer maps indicates a tendency for the perturbation (low pressure trough) to be long and of small amplitude when the index is large, and to be short and of large amplitude when the index is small. A relative eastward motion of the perturbation with larger zonal index is predicated on the assumption of constant dimensions of the perturbation. An attempt might be made to correlate the observed velocity of the perturbation with that computed from the formula (see p. 35), but further complications are introduced by the fact that frequently the perturbation does not lie well enough within the field of observation for accurate determination of its dimensions, and that its orientation is usually such that its latitudinal amplitude is not the same on either side of the crest. Thus it appears that at least until data for a more extended 3-kilometer map are available, theoretical considerations will not be of much help in forecasting the movement of this important feature of the pressure field aloft.

3. *Correlation of temperature anomalies in the eastern U. S. with the mean movement of the atmosphere aloft*

The orientation of the isobars on the weekly mean 3-kilometer pressure map in winter is closely associated with the current mean temperature anomalies, particularly in the northeastern U. S. This depends on the position of the upper level pressure trough and the northward or southward movement of air in the mean. Figures 30 and 31 illustrate conditions typical of cold and warm weeks, respectively, in the eastern U. S. The relationship is much weaker in the middle west because there, even in winter, relatively warm air from the Pacific, instead of cold continental air, frequently moves in from the northwest. But in the east it is exceptional, in winter, for warm air to be brought in from the northwest, or cold air from the southwest. However, when an abnormal temperature


0.7
0.6
0.5
0.4
0.3
LOW
0.3
0.4
0.5
0.6
0.7
0.8
HIGH
1.0
0.9
0.8
3 km. 5-Day Mean Pressure Map
March 19 - 23, 1939
Temperature Departure, Northeastern Forecast Section, -9°6 F


0.6
0.5
0.4
0.3
0.2
0.7
LOW
0.2
0.3
0.4
0.5
0.6
0.7
0.8
0.9
0.8
0.9
HIGH
3 km. 5-Day Mean Pressure Map
February 26 - March 2, 1939
Temperature Departure, Northeastern Forecast Section, +5°6 F

FIGS. 30, 31.

distribution makes such transport possible, then the usual relationship between the orientation of the isobars aloft and the current temperature distribution is correspondingly modified.

It is difficult to formulate any index of a general synoptic relationship of this kind that can be specific enough for statistical correlation and at the same time comprehensive enough to express the synoptic conditions.

One such attempt at a correlation was made. A rough index of northward or southward mean weekly air transport at 40°N 80°W was obtained simply by multiplying the pressure difference at the points (40°N 75°W–40°N 80°W) by the difference (40°N 80°W–45°N 80°W) on the weekly mean 3-kilometer maps. This product is negative for southward gradient transport (cold conditions) and positive for northward gradient transport (warm conditions). This weekly mean transport index was correlated with the weekly temperature departure from normal in the northeastern forecast section (New York State). The correlation coefficient was found to be +0.76 for the colder half of 1938–1939 (25 weeks) and +0.29 for 1939–40 (18 weeks). It seems probable that the 1938–39 coefficient is larger than would be found in the long run, but that the 1939–40 coefficient is considerably too low. The small coefficient for 1939–40 is believed a consequence of the persistent condition of extreme cold in the south, and comparatively mild cold in the north, during much of this past winter. A result of this condition was that some of the coldest weather to reach the northeast this winter came not from the northwest, as usual, but from the southwest. This greatly weakened the normal correlation between temperature anomaly in the northeast and the latitudinal component of the mean air transport.

The value of this type of correlation in the five-day forecasting lies in the fact that the 3-kilometer pressure map is much more conservative in its large features and consistent in its changes from day to day and from week to week than is the surface pressure map. Hence, significant changes in the general flow pattern can be better followed and forecast on this map than on the surface map. Therefore, it is of considerable practical assistance in the forecasting to be able to anticipate the surface conditions which should accompany given changes on the 3-kilometer map.

The effect of the northward or southward component of the mean air transport aloft can be seen in the surface temperature anomalies in the U. S. even for periods as long as a month, by plotting the monthly resultant winds aloft and comparing them with the surface temperature anomalies. However, when such long periods are investigated, it is found that the variations in the resultant winds are small for the same month from year to year, or from month to month, and there is little that can serve as a basis for the extrapolation of past conditions into the future. It follows that such comparisons have no forecasting value.

4. *Winter precipitation in the U. S. as related to upper air conditions in North America*

It has long been recognized that winter time precipitation in the southeastern United States for the most part falls irregularly and in the form of convective showers, even though the precipitation may be associated with a warm front. The physical reason for this characteristic is that the rain-producing air masses from the Gulf of Mexico are generally conditionally and frequently convectively unstable. A small amount of lifting over continental polar air masses is generally sufficient to liberate the energy of convection.

One of the consequences of this convective type of precipitation is that in any given storm the total amounts of precipitation are apt to vary considerably from station to station. The amounts recorded depend on whether or not the heavier convective showers pass over the station. It must not be inferred, however, that the winter precipitation of the southeastern section of the country is as localized as summer showers. If a sufficient number of stations is considered, it is possible to obtain variations in precipitation from one period to another which are distinctly associated with air mass and frontal distributions. The first points to be considered, then are:—(1) How is the potential instability of the Tropical air masses established? and (2) Are there any factors which might indicate the degree of instability in the period when the tropical air is being manufactured?

The correct answer to these two questions should go a long way toward solving the forecasting of general precipitation over the eastern United States in winter.

The following study was carried on entirely with data from the winters of 1938–39 and 1939–40. Before this time aerological soundings were made by airplane (instead of radiosonde) and observations in the free air were insufficient to carry on a careful survey. Since the study was in large part suggested by empirical findings gleaned from the daily analysis of upper air data of 1938–39, it is not surprising to find that the methods suggested on the basis of that material were less successful in their application to 1939–1940. Nevertheless, it is believed that the results are sufficiently encouraging to warrant usage, and their success during the last two winters suggests that at least we are on the right track to a more adequate solution of the problem.

It has generally been assumed that most of the tropical maritime air responsible for the winter-time precipitation in the eastern United States originates in the sub-tropical high-pressure cell of the North Atlantic. According to Bjerknes* this tropical air, in its trajectory around the south side of the North Atlantic High, ascends and converges as it moves westward, so that a state of conditional instability is established. Bjerknes points out that the regions of most frequent shower-activity appear to be located in the western (particularly southwestern) portion of the subtropical anticyclonic cells, while the eastern portions, where descent of air is taking place, are characterized by deficient precipitation.

Rossby and Weightman,† in a study of the synoptic weather situation for the period February 16–19, 1926 were probably the first to point out the importance of conditional instability for warm front precipitation over the southeastern United States. But they attempted to explain the conditional instability on a different basis than did Bjerknes some years later. Instead of ascribing it to thermodynamic processes taking place in the sub-tropical source region, they suggested that it was due to veering of wind with elevation so that progressively colder air from westerly regions is carried over warm southerly air. This idea, also advanced by Humphreys for the explanation of thunderstorms in the southeastern quadrants of cyclones, has not received the attention it deserves, perhaps because of over-emphasis of the horizontal homogeneity of air masses.

The daily analysis of aerological data over the United States has shown that many different vertical distributions of temperature and moisture may be observed in tropical maritime air masses, while at the earth's surface, these air masses have essentially the same properties from one occurrence to another.

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* Bjerknes, J., *Physikalische Hydrodynamik*, J. Springer, Berlin, 1933.

† "Application of the Polar Front Theory to a Series of American Weather Maps," *Monthly Weather Review*, Vol. 54, No. 12, 1926.

The fact that the tropical air masses in the United States differ so widely in their properties in the free atmosphere suggests *per se* that their origin cannot be attributed solely to one source region.

Some clues as to the differences in the source properties of so-called tropical air have been suggested by the daily analysis of Northern Hemisphere charts and upper-air soundings over the United States. Numerous cases have been observed in which the polar front has moved far out over the North Atlantic, polar air drains into the trade-wind belt, and frontogenesis occurs just off the Gulf coast or off the coast of the Carolinas. Such frontogenesis normally takes place between the polar air masses which have had a trajectory over the water and those which have followed a path which is mainly over the land. During the winter of 1938–39 cases of such frontogenesis usually occurred following an outbreak of especially cold air at intermediate levels (about 4 kilometers) over the northeastern United States.

On the basis of past experience with the Northern Hemisphere charts, it seems probable that tropical maritime air masses can be manufactured within a few days by the intense convective activity that must set in as deep cold polar air moves over the warm waters of the Gulf Stream and Gulf of Mexico. (When an especially cold polar outbreak occurs at Boston, one can see evidence of such convection in the Cunb clouds over the ocean within 30 miles.) This type of air mass is unstable and showery, and is associated with lower than average "tropical air" temperatures at upper levels. An extreme example of such a development is afforded by the period February 22 to March 1, 1939, in which air leaving the Continent at upper levels over Lakehurst and Washington was extremely cold (−38° minimum at 600 mb over Washington.) Shortly after this air mass had left the Continent, extensive precipitation occurred over the eastern United States, and the heaviest 5-day totals of precipitation for the winter season were recorded. The front between tropical and polar air is usually steep in such cases, and the temperature-difference between polar and tropical air is comparatively small. Soundings at Nashville, for example, frequently show that the tropical air is a few degrees colder than normal tropical air temperature. The fronts, therefore, are mainly moisture fronts, and not much ascent of moist air is required to produce precipitation. Since the air is unstable by virtue of its life history, the convection, once started, is assisted by the vertical temperature distribution.

It may be concluded, then, that some of our tropical maritime currents are manufactured by convection in deep polar air masses moving over the warm seas. It is interesting to note that in summer the same trajectory of polar air eventually leads to our dry and warm air (T_s) currents.

The question next arises as to what temperatures aloft are necessary to allow convection over the Atlantic to penetrate a fairly deep stratum. If the 600-mb surface is arbitrarily chosen as that level to which convection must proceed to form a significantly deep moist air mass, then it is possible to estimate the temperature at this level that must be exceeded in order for air following a trajectory over the Gulf Stream to permit convection through this level. Using the value of mean winter pressure in this area given by Shaw,* and the temperatures of the surface-water in the Gulf Stream by Slocum,† and assuming a dry adiabatic lapse rate, this "critical" temperature is found to be −18°C.

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* "Manual of Meteorology," Vol. 2, Comparative Meteorology. 1936.

† "The Normal Temperature Distribution of the Surface Water of the Western North Atlantic Ocean," *Monthly Weather Review*, Vol. 66, No. 2, 1938.

TABLE V

1938–1939

PERIOD	① PRECIP. 12 S.E. STATIONS	② OBSERVED DN	③ MIN. TEMP. AT 600 mb. AT LP IN PRECEDING 5 DAYS	④ MIN. TEMP. AT 600 mb. AT EO IN PRECEDING 5 DAYS	⑤	⑥ FORECASTS FROM LP	⑦ FORECASTS FROM EO	⑧ FINAL FORECAST
Dec. 1– 5	11.10	+						
Dec. 6– 10	4.25	−	−16	−10	−26	−	−	−
Dec. 11–15	.19	−	−18	− 6	−24	−	−	−
Dec. 16–20	1.15	−	−16	−10	−26	−	−	−
Dec. 21–25	5.44	−	−21	− 8	−29	+	−	−
Dec. 26–30	10.90	+	−25	−14	−39	+	+	+
Dec. 31–Jan. 4	6.13	−	−16	− 8	−24	−	−	−
Jan. 5– 9	2.35	−	−19	− 8	−27	+	−	−
Jan. 10–14	11.40	+	−17	−15	−32	−	+	−
Jan. 15–19	6.32	−	−14	−14	−28	−	+	−
Jan. 20–24	4.30	−	−20	−10	−30	+	−	−
Jan. 25–29	13.99	+	−24	− 0	−31	+	−	−
Jan. 30–Feb. 3	16.66	+	−29	−16	−45	+	+	+
Feb. 4– 8	8.26	−	−17	−20	−37	−	+	−
Feb. 9–13	12.14	+	−19	−12	−31	+	+	+
Feb. 14–18	11.10	+	−20	−16	−36	+	+	+
Feb. 19–23	8.44	+	−21	−14	−35	+	+	+
Feb. 24–28	31.56	+	−38	−16	−54	+	+	+
Mar. 1–5	13.14	+	−18	−19	−37	−	+	−
Mar. 6–10	3.33	−	−16	− 9	−27	−	−	−
Mar. 11–15	7.30	−	−16	− 7	−23	−	−	−

1939–1940

PERIOD	① PRECIP. 12 S.E. STATIONS	② OBSERVED DN	③ MIN. TEMP. AT 600 mb. AT LP IN PRECEDING 5 DAYS	④ MIN. TEMP. AT 600 mb. AT EO IN PRECEDING 5 DAYS	⑤	⑥ FORECASTS FROM LP	⑦ FORECASTS FROM EO	⑧ FINAL FORECAST
Dec. 1– 5	3.18	−						
Dec. 8–10	0.60	−	−20	− 6	−26	+	−	−
Dec. 11–15	2.40	−	−21	− 6	−27	+	−	−
Dec. 16–20	5.28	−	−24	− 5	−29	+	−	−
Dec. 21–25	6.65	−	−17	− 8	−25	−	−	−
Dec. 26–30	10.24	+	−22	−11	−33	+	+	+
Dec. 31–Jan. 4	1.24	−	−21	−15	−36	+	+	+
Jan. 5– 9	6.89	−	−30	− 9	−39	+	−	−
Jan. 10–14	17.17	+	−26	−16	−42	+	+	+
Jan. 15–19	.57	−	−17	−12	−29	−	+	−
Jan. 20–24	8.74	+	−27	−11	−38	+	+	+
Jan. 25–29	.07	−	−28	−14	−42	+	+	+
Jan. 30–Feb. 3	.22	−	−32	−10	−42	+	−	−
Feb. 4– 8	10.88	+	−23	−11	−34	+	+	+
Feb. 9–13	14.62	+	−21	−17	−38	+	+	+
Feb. 14–18	17.15	+	−15	−18	−33	−	+	−
Feb. 19–23	1.88	−	−14	−17	−31	−	+	−
Feb. 24–28	4.83	−	−20	−10	−30	+	−	−
Feb. 29–Mar. 4	4.16	−	−25	− 9	−34	+	−	−
Mar. 5– 9	1.88	−	−18	−11	−29	−	+	−
Mar. 10–14	18.08	+	−17	−11	−28	−	+	−

Normal Precip. Sum for 12 SE Stations

Dec.	7.58
Jan.	7.85
Feb.	8.39
Mar.	8.52

On the preliminary assumption that *all* air leaving the Continent over Lakehurst undergoes the anticyclonic motion necessary to return it to the southeast coast, the following forecast for the succeeding 5-day period in the southeastern United States was tried: If the temperature at the 600-mb level over Lakehurst (or Washington in the absence of Lakehurst's sounding) drops below −18°C in any one day (or more) of a five-day period, the succeeding five-day period should be characterized by above-normal precipitation, otherwise by sub-normal precipitation. In the winter of 1938–1939 (December 5 to March 10) the application of this simple rule would lead to ten forecasts of greater than normal precipitation in the southeast and ten forecasts of less than normal. These forecasts are shown in column 6 of Table V. Upon verifying the forecasts on the basis of the total precipitation at 12 stations in the southeastern United States* (column 1) there was found fifteen forecasts correct and five incorrect. For the same period in 1939–1940 (Table V), there were nine correct and eleven incorrect. A rough measure of the degree to which the minimum temperature at the 600-mb level over Lakehurst affects the precipitation of the southeast is also shown by the correlation coefficients between the former element (Lakehurst 600-mb minimum temperature) and the total precipitation observed in the following five days at the 12 southeastern stations. These coefficients were:

1938–39 $r = -0.58$
1939–40 $r = +0.14$

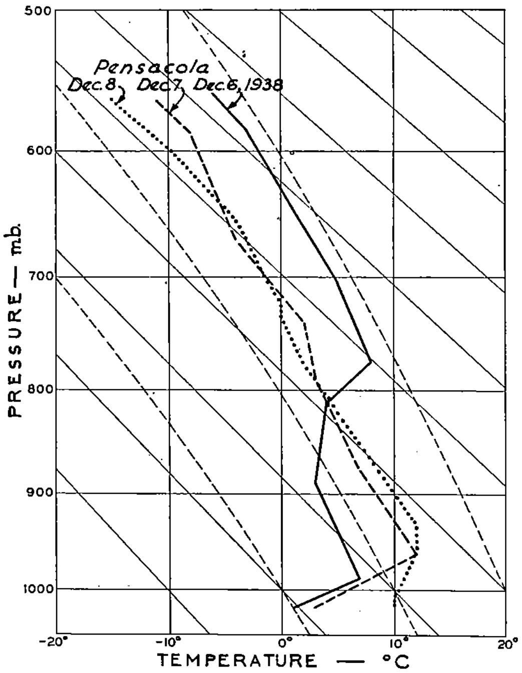

FIG. 32. Steepening lapse rates observed over Pensacola, Florida.

The correlation between the minimum temperature at the 600-mb level over Washington, D. C., during one five-day period and the departure from normal of precipitation during the following five-day period at the three stations which verify the southeast forecast section (Birmingham, Atlanta, and Chattanooga) for the period October 23, 1938, to April 6, 1939 was found to be −0.70.

The above evidence seems to indicate that the nature of the upper level polar air leaving the continent has some influence on the following five-day precipitation in the southeast. This was much more effective in the winter of 1938–39 than it was in 1939–40, and this difference may be associated with the fact that the Bermuda High was well developed during much of the winter of 1938–39, while it was largely absent during most of the winter of 1939–40. Thus it is possible that in the latter year the polar air, for the most part, moved off over the Atlantic and seldom was forced to take a trajectory over the Gulf, to return to the southeastern United States.

Another important consideration is that of the consequences produced by the establishment of an unstable air mass stratification in the southeast.

As an illustration of the importance of this type of action, the weather situation of December 6 to 8, 1938, has been chosen. In Fig. 32 are shown the soundings made during

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* These stations are Indianapolis, Elkins, Washington, Nashville, Asheville, Raleigh, Hatteras, Birmingham, Atlanta, Charleston, Mobile, Thomasville.

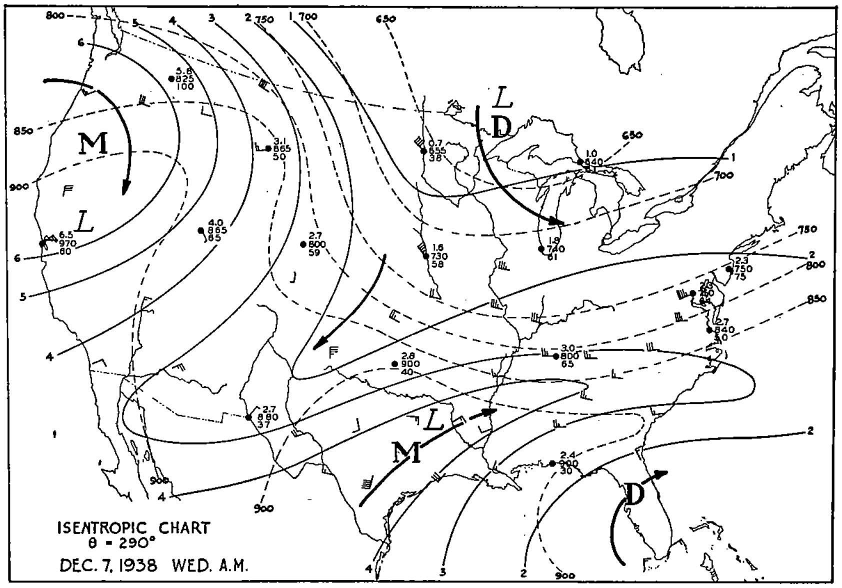

Fig. 33. Isentropic chart for $\theta=290°$, December 7, 1938. Solid lines refer to mixing ratio (g/kg), broken lines to pressure in mb. Data entered beside aerological stations in the order top to bottom, mixing ratio, pressure, relative humidity.

this period at Pensacola, Florida. In the entire column of air from the surface to the top of the ascents a remarkable change in the vertical stability takes place between December 6 and 7. At levels above 810 mb there is a drop of temperature amounting roughly to 6°C, while below this level there have been equally large *increases* in tempera-

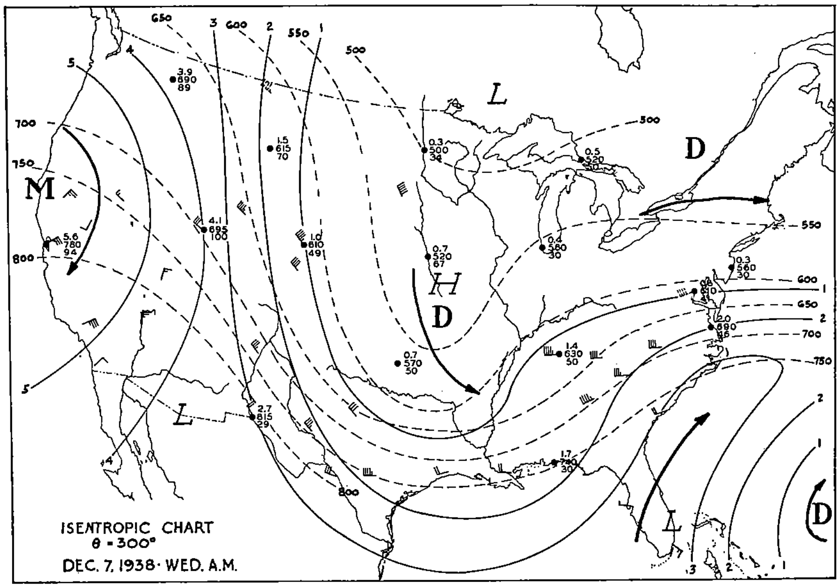

Fig. 34. Isentropic chart for $\theta=300°$, December 7, 1938.

ture. This change in the lapse-rate would indicate that cold air has been brought in aloft while warmer air has been transported in the lower layers. This "differential" advection is best shown by the isentropic charts (Figs. 33 and 34) for two surfaces of constant potential temperature (290° and 300°). In the lower isentropic surface (Fig. 33) a tongue of moist and warm air is observed flowing northeastward over the east Gulf States. Aloft, however, the axis of this moist tongue is shifted eastward, and has been displaced by the tongue of cold and dry air from the west.

The store of potential energy thus built up finally breaks down on December 8 when thundershowers are observed in northern Florida.

In the case treated above, there seems to be some evidence that the showers produced on December 8 at Pensacola and Apalachicola, Florida, were the simple result of heating the surface-air to temperatures sufficiently high (66° at Apalachicola) to release the stored-up energy in the conditionally unstable air over this area. At least, an energy-diagram for the ascent on December 8 shows that the observed maximum temperatures were sufficient to release the conditional instability aloft.

In an attempt to make use of this theory in forecasting the departure from normal of the precipitation for five-day periods in the southeastern United States, especial attention has been devoted to the temperatures at upper levels further west, notably at El Paso, Texas.

As a first approximation the air over El Paso, at the 600-mb level, is assumed to take such a trajectory that it always moves across the south eastern portion of the United States. The next step is to choose a "critical" temperature which if exceeded will permit convection to reach the 600-mb level as the air moves over the southeastern states. This critical temperature is obviously not as easily fixed as for polar air moving over the Gulf Stream, for the lower air in this case may acquire different temperatures under the influence of insolation. The normal temperature for the 600-mb level over El Paso for the winter season is around -5°C. As a reasonable critical value -10°C has been chosen.

Forecasts based on this critical value were made for the southeast (the same 12 stations) in the same manner as those based on the Lakehurst forecasts. That is, if the minimum temperature at the 600-mb level over El Paso fell below the critical value (-10°C) on any one of five days preceding the day the forecast was made, then the total precipitation for the following five-day period should be above normal, otherwise it should be below normal. These forecasts are given in column 7 of Table V. In the winter of 1938–39 (from December 6 to March 15) there were ten forecasts for above normal precipitation and ten for below normal. Of the ten above normal, two failed, and of the ten for below normal one failed. Thus there were seventeen forecasts correct and three incorrect. In 1939–40 there were, by this method, twelve forecasts of above normal precipitation, seven of which verified, and eight forecasts of less than normal precipitation, all of which verified. The correlation coefficient between the El Paso minimum temperature at 600 mb and the following five-day precipitation totals at the 12 southeastern stations was -0.53 in the winter of 1938–39 (for the periods given in Table V) and -0.38 in 1939–40. This factor, then, as well as the eastern polar outbreak factor, appears to be significant.

Indeed, it appears reasonable that both factors may act simultaneously, for the creation of conditional and convective instability requires two conditions: (1) the advection of warm, moist southerly air, and (2) the advection of cold air from the west above this warm stratum. The second factor is, to some degree, taken care of in the El Paso criterion, which enjoyed appreciable success in both winters. The former condition may be associated with the coldness of polar air leaving the continent. In the winter of 1938–39, for example, there was found a correlation of −0.56 between the minimum 600-mb level temperature observed over Lakehurst and the mean surface pressure observed at Bermuda in the following five days. In 1939–40, however, a similar correlation based on seven day means of pressure at Bermuda was found to be insignificant. At least in the winter of 1938–39, it appeared possible that the advection of especially cold air over the eastern United States stimulated a subsequent intensification of the Bermuda High. This increased high pressure may have been responsible for the warm, moist currents in the southeast which assist in the creation of instability as the cold air from the west overruns them.

Assuming that some such combination of both factors creates the instability of the southeast, the following combination forecast was made: When forecasts based on the critical temperatures at Lakehurst and El Paso (columns 6 and 7 of Table V) agree in sign, there will be a distinct departure of precipitation in the indicated direction. For example, if both criteria suggest negative departures, precipitation will be subnormal; if both criteria suggest positive departures, above normal. If the criteria disagree, somewhat less than normal precipitation may be forecast. (Because normal precipitation occurs much less than half the time, and because the heavy amounts presumably fall only when both critical temperatures are exceeded.)

This final forecast is shown in column 8 of Table V. In the winter of 1938–39 there were six forecasts of above normal precipitation for the southeastern 12 stations made, all of which verified, and there were 14 forecasts of subnormal precipitation, only three of which failed. In the winter of 1939–40 there were seven forecasts of above normal precipitation made, five of which verified, and thirteen forecasts of below normal precipitation made, eleven of which verified. In all, then, there were 33 out of 40 forecasts correct.

The correlation coefficient between the sum of the Lakehurst and El Paso minimum temperatures at 600 mb and the following five-day totals of precipitation at the twelve southeastern stations was −0.71 in 1938–39, and −0.14 in 1939–40. For all the 40 periods in these two years it was −0.52.

There were other correlations attempted which might throw some light on the question of five-day precipitation totals. One of these was between the departure from normal of the mean lapse rates in the three forecast districts (taken from the 600-mb to the 900-mb surface) with the five-day precipitation totals for the same period. The period for this correlation extended from October 23, 1938, to April 6, 1939. The correlation coefficient was found to be insignificantly small. The reason for this negligibly small coefficient may be that on one or two days in which all the precipitation falls the lapse rate may be quite steep, yet be overbalanced in the means by stable conditions. Then again, the layer 900 to 600 mb may contain a frontal discontinuity, in which case the lapse rate would not be representative of the instability above the front.

A correlation was made of the departure from normal of the five-day mean temperature at the 600-mb surface at the three forecast districts and the departure from normal of the simultaneous five-day precipitation totals for the same period as the above. This coefficient was nearly zero, which shows the lack of effect of this factor alone in producing precipitation. Indeed the coldest air aloft is generally the dry Pc air masses, while the warmest air is also dry—the T_s air.

A correlation of the five-day mean temperature at 600 mb at El Paso and the simultaneous five-day precipitation totals in the southeast came out to be −0.42, indicating some of the correlation discussed in the above treatment of El Paso temperatures aloft.

Finally, a correlation was made between the departure from normal of the five-day precipitation in all three forecast districts (using the total sum) with the current departure from normal of the horizontal temperature gradient at 3 kilometers from Washington, D. C., to Omaha, Nebraska. The coefficient was −0.50, which strongly indicates that above normal precipitation in the eastern United States tends to occur when the east-west temperature gradient aloft (at about the 3 kilometer level) is steep—i.e., when the east is warm and the west cold. This may be interpreted as indicating the presence of an important frontal zone with strong concentration of solenoids in the eastern United States. Moreover, this is in agreement with the empirical result found from isentropic analysis that precipitation is chiefly confined to the left side of moist, warm tongues.

Thus far the precipitation has been discussed chiefly from the standpoint of the source of moisture and the origin of instability in the rain-producing air masses. It has been taken for granted, for example, that there is always available in winter enough frontal activity to release this instability in the form of precipitation. But it is obvious that the position and intensity of the polar front surface must be taken into account in order to apply the above reasoning in order to get the best results. This is indicated by the last correlation discussed in the preceding paragraph. For this reason the problem is closely allied with that of the temperature forecasts, and the same problems of the intensity and location of the centers of action arises. It is believed, however, that certain additional clues to the problem might be found in the further analysis of upper air data, particularly since the density of the aerological network has been greatly increased. The problem, when viewed from this angle, appears to lend itself especially well to a statistical type of research.

5. *Use of the American Aerological Data in the Study of the Summertime Temperature and Precipitation Anomalies in the United States*

a. *Introduction.* Although the great bulk of the work at M.I.T. on the five-day forecast project has been carried out during the school year, so that winter weather has received first attention, nevertheless the program has been carried through the summer seasons with results of considerable interest, as appears in the following.

Since July, 1934, there has been established a fairly close network of aerological sounding stations over the United States. With the data obtained from these soundings it has been possible to obtain a more accurate picture of the prevailing circulation. The nature of this observed upper air circulation, particularly in summer, was not anticipated before the upper air data were analysed along lines suggested by Rossby—i.e., by means of isentropic analysis.

More recently the method of isentropic analysis has been used to study the mean warm season circulation over the United States as portrayed by monthly mean values for the months May through September, 1934–39. The data from which these charts were prepared are published in the *Monthly Weather Review* in the form of tables giving the mean temperature, pressure, and relative humidity for standard levels at the aerological sounding stations. The method consists of making an adiabatic diagram, one for each station, for each "mean" sounding, and from these diagrams are computed the elevation, mixing ratio, and relative humidity at the prescribed isentropic surface chosen for study. In addition, the resultant winds, also published in the Review, are entered on the isentropic chart at the elevations determined by an analysis of the contour lines on the isentropic chart.

The drawing of these mean isentropic charts is definitely *not* a mechanical procedure, for the network of aerological stations is far too sparse to permit a unique mechanical solution. The *analysis*, therefore, consists in (1) applying a well-rounded theory developed by Rossby, and discussed in Section I; and (2) making use of certain indirect clues to the true flow pattern as offered by such surface features as departures from normal of temperature, sunshine, precipitation, etc. When these factors are taken into account, it becomes possible to construct analyses which have definite form and characteristics and which, in themselves, express a type of circulation. Extended experience with such analyses, including some independent checks, makes it appear that, even to considerable detail, they represent correctly the upper air conditions.

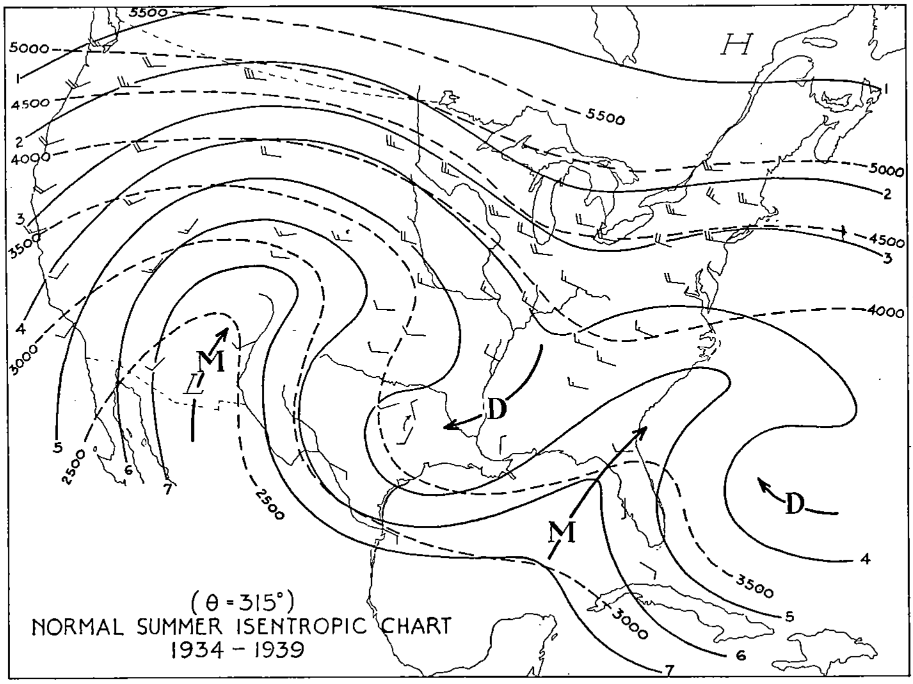

Fig. 35. Normal summer isentropic flow pattern. Solid lines refer to mixing ratio, in g/kg, broken lines to elevation in meters above sea level. Winds are normal resultants computed for a long series of pilot balloon observations, sometimes not coincident with period for which the aerological sounding data was obtained.

b. *The Normal Circulation.* By making use of the analysed monthly mean isentropic charts for the summer months June, July and August, 1934–1939, it was possible to obtain a fairly reasonable picture of the normal summertime circulation over the United States. This was done by averaging the interpolated values of mixing ratio and elevation for five degree intersections of latitude and longitude. The resultant values were then plotted, and the "normal" chart was uniquely defined by the large number of values.

There is, of course, some doubt as to the real normal state, since the number of summer months (17) from which the chart was constructed is, climatologically speaking, small. Moreover, during this period the temperature over the United States as a whole averaged warmer than normal. However, because certain characteristic features of the normal charts are also expressed in monthly charts where the departures of surface temperature are small, it seems reasonable to conclude that the real normal picture does not differ much from the one obtained from the data for the summer months 1934–39.

The outstanding feature of the normal summer pattern (Fig. 35) is the existence of two well-defined anticyclonic cells—one centered over western Texas, the other somewhere off the southeastern coastal states. The fact that *two* eddies appear in such geographically different places seems to preclude the possibility that these eddies are entirely thermally produced by continental heating. It is probable, however that the western cell is somewhat intensified in this manner. Some light on the nature of these eddies is furnished by north-south atmospheric cross sections, a typical one for the summer season being reproduced in Fig. 36. This section brings out the well-known fact that over North America the principal summertime extra-tropical front is generally found in the vicinity of the Canadian-U. S. border. Most of the time the United States is south of this front and is within what appears as a thermally homogeneous air mass. In spite of the zonal homogeneity of temperature and the consequent lack of solenoids to generate kinetic energy, there is observed a prevailing eastward flow of the tropical air. According to Rossby the eastward current in this homogeneous air is maintained by frictional stresses from the much stronger westerly current to the north, this energy continually being dissipated in the form of eddies on the southern edge of the current. If such eddies are maintained in more or less fixed locations over a sufficiently long period of time, the mean chart will display an eddy pattern in the moisture lines and in the upper air resultant winds. The mean isentropic charts for five-day periods during the summer months reveal such eddies a large part of the time, as do the individual summer monthly mean charts.

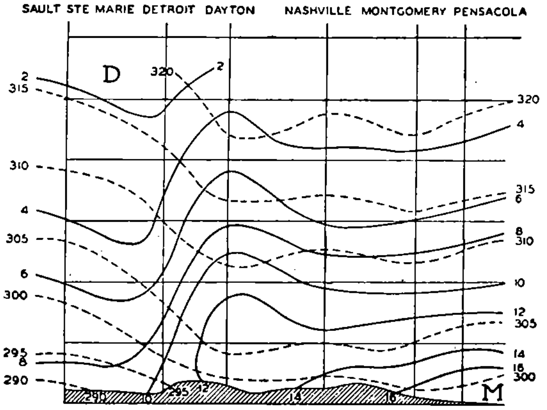

FIG. 36. Mean north-south cross section, Sault Ste. Marie, Michigan, to Pensacola, Florida, for August, 1936. Solid lines refer to mixing ratio (g/kg), broken lines to potential temperature (°A).

In all the mean summer isentropic charts studied there has been a moist tongue projecting from Northern Mexico with its axis located over Arizona or New Mexico. The moist tongue invariably curves off to the east after moving some distance northward, but the northward displacement before the current has curved around into an eastward flow appears to vary considerably from month to month. The origin of this moist tongue is the subtropical easterlies which, on ascending the high Mexican Plateau, carry moisture and heat to high levels. This is substantiated by the fact that the rainy season for central and eastern Mexico is in summer and that the rainfall maximum in southwestern Texas, New Mexico, and eastern Arizona is also in summer.

The fact that the principal moist tongue always seems to make its appearance in this locality does not, however, explain why it should proceed to move northwards and gradually be deflected eastward and then southward, making a gradual anticyclonic sweep. Converging with this tongue in a spiraliform fashion into the center of the eddy is a dry current emanating from Canada, and normally projecting from the region of the Great Lakes. It has been suggested by Wexler* that this preferred site for the extensive anticyclonic eddy is largely due to topography and results from the field of solenoids established on the side of the warm Rocky Mountain region as well as from the field of solenoids to the north. The former field starts the moist tongue on the northward leg of its journey, and as this field combines with the frictionally driven eastward motion, the moist tongue swerves to the east. Given such a geographical distribution of solenoids it appears that a "keystone" eddy is normally established just east of the Southern Rockies, and other anticyclonic eddies are formed at some distance downstream. The pattern thus envisioned is similar to the right-hand side of a Karman vortex street.

The eastern eddy (Fig. 35) usually has the axis of its moist tongue over Florida, and this region is associated with a secondary maximum of thunderstorm and shower activity.

The general relationship between the positions of the eddies and the climatic distribution of precipitation was brought to light by Wexler, who showed by the analysis of 50 years of rainfall records that the percentage of annual rainfall occurring in August seemed to show a rainfall pattern suggested by certain mean isentropic charts constructed for the August months, 1934–1937.

The now well-established fact that summer rainfall is confined chiefly to moist tongues, while the dry tongues are relatively free of showers, is closely tied up with their characteristic vertical distributions of temperature and moisture. Consider two models: one, an atmosphere in which the high moisture content is limited to a shallow surface layer (about 2 km); and in which there is established between the dry and moist air a dry-type discontinuity, usually an inversion of stable layers; the other a deep, moist flow, generally free of discontinuities. In the first model the stable layer hinders growth and progress of convection currents from below and also encourages intense lateral mixing so that moist currents entering the layer from below are soon robbed of their moisture by mixing with dry air. Moreover, the upward momentum of convective columns is disminished by this mixing. The resultant effect is to suppress convection from below and to raise the condensation level of the rising air to a level to which convection is not likely to reach. Even if convection does reach this high level the convective rain must fall through a thick layer of dry air where it is readily evaporated. In the case of deep moist currents, no stable dry type discontinuity is formed, so that the convection, once started, can penetrate to great heights without the damping effects of lateral mixing. Showers are therefore found chiefly in the moist tongues.

It is also observed that the heaviest rainfall occurs on the left side of the moist tongues. Since the contour lines have an appreciable upslope gradient here, it appears that the lifting effect must also be considered in the distribution of summer rainfall. The upslope effect becomes increasingly important, as the latitude increases.

c. *The Pressure Distribution at the Surface and Aloft.* Since the anticyclonic eddies are observed on isentropic surfaces, it is not suprrising to find that resultant winds at fixed

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* "Observed Transverse Circulations in the Atmosphere, and their Climatological Implications," Doctor's Thesis at Mass. Inst. Tech., (unpublished).

levels also are closed anticyclonic wind systems. As the anticyclonic eddies are believed to be frictionally driven from the strong westerlies to the north, there takes place from day to day a general banking of the westerlies and convergence of air into the eddies. The convergence produces an accumulation of air above fixed levels, which in turn creates an anticyclonic pressure distribution.

In order to study the circulation at fixed levels there were obtained from the Weather Bureau records of the normal resultant winds at many stations over the United States. These normals were based on data from thousands of pilot balloon observations made during the last 20 years, although most of these data are for periods of about six years. These resultant winds offer a good picture of the normal circulation and pressure distribution observed at their respective levels over the United States in summer (June, July and August) as shown in Fig. 37. At 750 m the resultant winds show the characteristics already well known from the map of normal sea level pressure, especially the south and southwesterly flow of air (generally of tropical origin) over the eastern part of the country. It will be noted that, just as on the mean sea level pressure map, there is no anticyclonic cell corresponding to the western anticyclonic eddy of the normal isentropic chart. Rather, there is an extensive intrusion over the continent of the Bermuda High. At 1500 m a similar pattern obtains, and still the western anticyclonic cell fails to appear. The westerlies in the northern part of the country have increased in strength. At 3000 m the anticyclonic cell definitely appears as a closed circulation whose center is located over Louisiana, and at 4000 m the center of the cell is displaced westward over Texas, roughly in the same location as the western anticyclonic cell of the normal

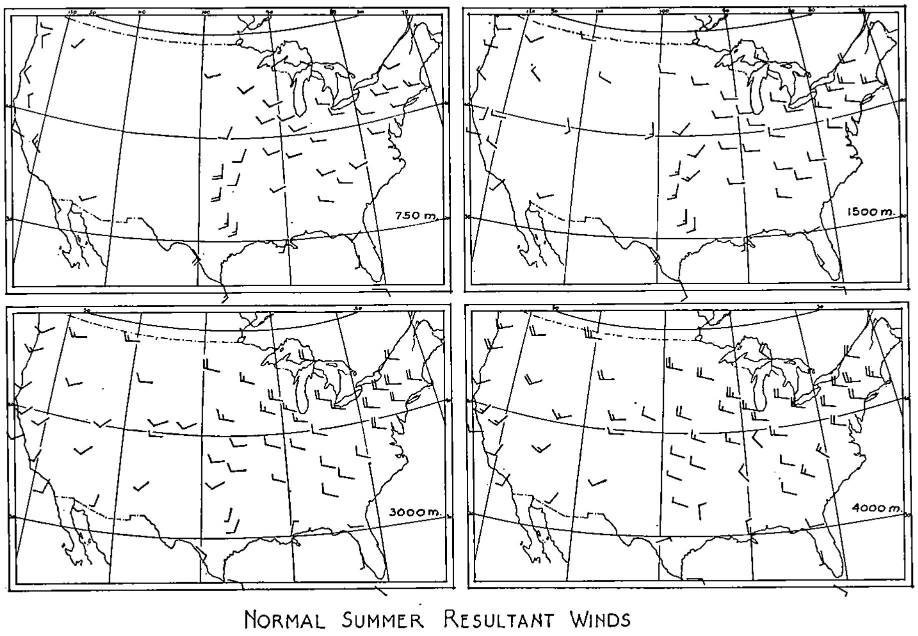

FIG. 37. Normal summer resultant winds at 750 m, 1500 m, 3000 m and 4000 m.

isentropic chart (O=315°). At the same time the eastern cell continues up to high levels, as shown by the winds over Florida.

The question naturally arises as to why the western anticyclone is confined entirely to upper levels, there being no reflection of it whatsoever in the surface pressure distribution. The structure of the anticyclonic cell as seen from the isentropic analysis consists chiefly of an extensive mass of dry air—air which has come from the westerlies over Canada and which subsides as it travels southward above the surface shallow layer of moist Tropical Gulf air. The showerless, and often cloudless, type of weather is therefore present over the midwest when this upper level anticyclone is well developed so that the intense insolation heats the lower layers of air to such an extent that the contribution of the upper level high to the surface pressure is completely nullified. More will be said in regard to this subject later.

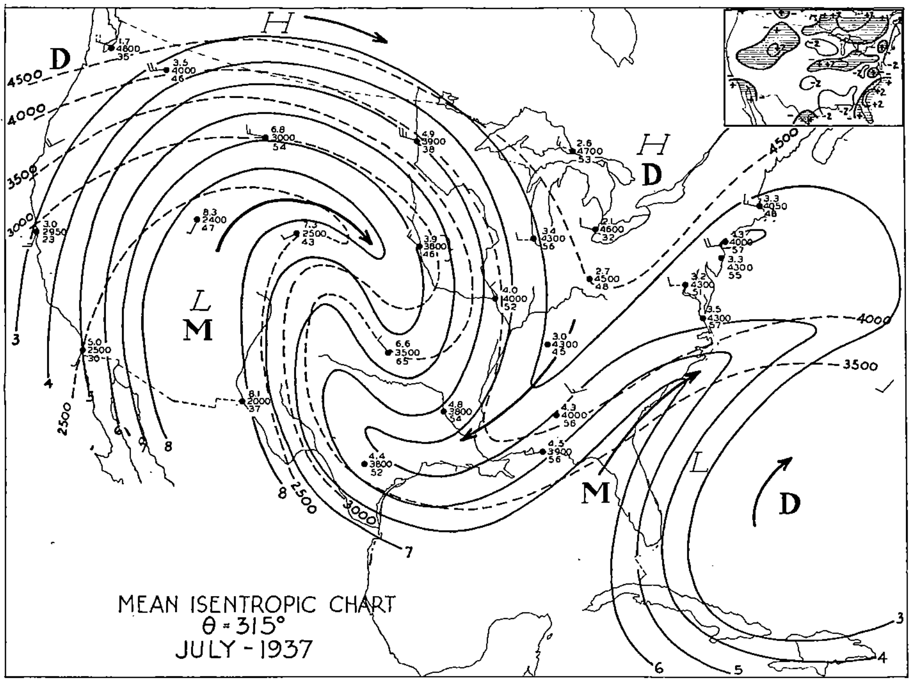

Fig. 38. Mean isentropic chart for July, 1937. (Inset—departure of precipitation from normal expressed in inches.)

d. *Variations of the Monthly Mean Flow Patterns from the Normal.* While there is a similarity of mean monthly isentropic charts for different summer months and from year to year, it should not be inferred that the moisture patterns are always the same. Considerable deviations from the mean or normal state are always associated with considerable anomalies of rainfall, temperature, amount of sunshine, etc. Thus while a fairly normal distribution of precipitation occurred in July, 1937, and for this month the mean isentropic chart (Fig. 38) appears quite similar to the normal chart (Fig. 35) (p. 71), one may contrast the month of August, 1936, when the entire midwest was very hot and dry and when the mean isentropic chart (Fig. 39) showed an abnormally large anticyclonic eddy over the United States, rather than the double-cellular pattern usually present

From a study of the mean monthly isentropic charts it may be readily concluded that drought months in the midwest are associated with especially well developed western anticyclonic eddies. This fact had indeed been pointed out in 1933 by Reed,* who studied the upper level anticyclones by means of monthly resultant pilot balloon winds. The true nature of these cells, however, is more readily understood from the analysis of the mean isentropic charts. Thus the correlation between the presence and development of the western anticyclonic cell and the associated precipitation is chiefly a result of the advection of subsiding dry Canadian air into the anticyclonic cell. This subsiding dry air, moving over the shallow layer of moist Tropical Gulf air, presents the ideal conditions for suppressing shower activity, as explained in the foregoing section (See p. 73). The better developed the western anticyclonic cell, the stronger and more extensive becomes this dry air flow. As the dry air envelops the cell it also helps to strengthen its circulation. This is accomplished by means of the intense heating which takes place in the region below the dry air cast off from the westerlies, and this dry air produces ideal conditions for intense insolational heating, which in turn strengthens the anticyclonic circulation.

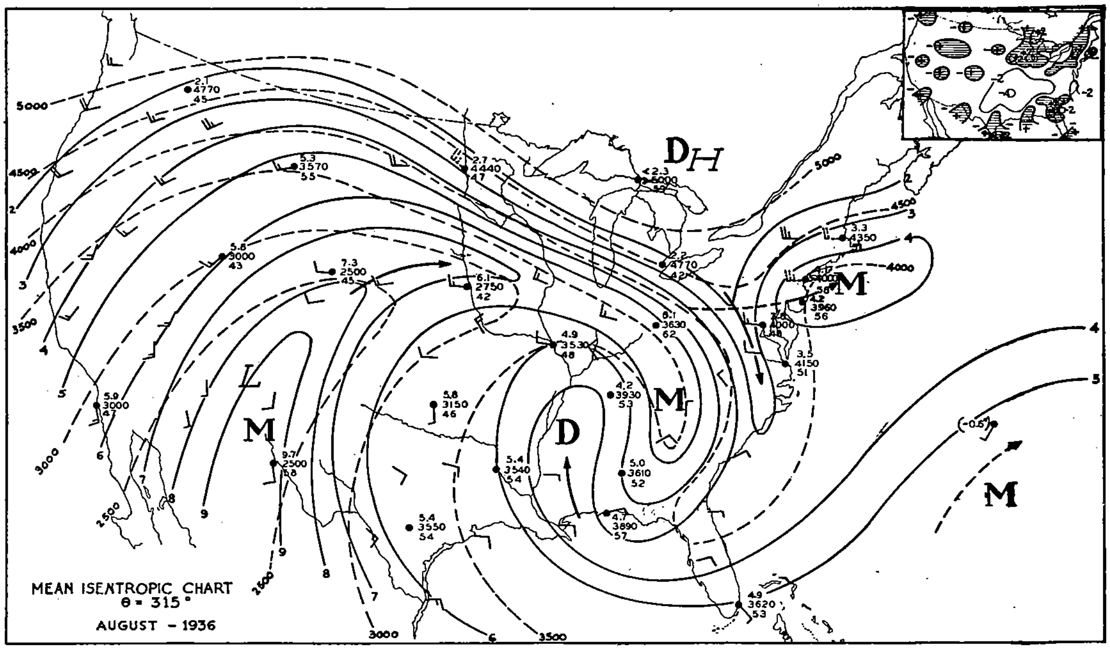

FIG. 39. Mean isentropic chart for August, 1936. (Inset—departure of precipitation from normal expressed in inches.)

An illustration of these upper level effects may be had by comparing the mean values of the soundings made over Oklahoma City in August, 1936 (a hot, dry month in the midwest with temperature 6.6°F above normal and precipitation −2.71 inches below

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* *Monthly Weather Review*, Vol. 61, 1933, page 321.

normal in Oklahoma), and in August, 1937 (when Oklahoma was only 2.7°F above normal and had a precipitation deficit of only .34 inches).

The upper air data for these two months as well as the August months of 1934, 1935, 1938 and 1939 are shown in Fig. 40. The solid lines are lines of equal mean specific humidity, and the broken lines are mean isentropic surfaces. The striking characteristics of August, 1936, as compared with other Augusts are:

1. The exceptionally dry air on all isentropic sheets from the surface to 600 mb.
2. The marked warmth of the lower strata of the atmosphere (up to about 850 mb).
3. The fact that aloft there is comparatively little temperature difference from one August to another, and that the lapse rate at levels above 800 mb is comparatively steep in 1936.

These characteristics of August 1936 (Fig. 40) are probably characteristic of all severe summer drought months in Oklahoma, and these conditions appear to be in complete agreement with an isentropic analysis of the data. The extreme relative dryness of 1936 is attributed to the exceptionally well developed anticyclonic eddy into which the Polar Canadian air is driven. In this particular month the eddy was so extensive that a good portion of this polar air was forced to make an extended anticyclonic sweep over the western North Atlantic before invading the Mississippi Valley (Fig. 39). Thus it had ample time to subside undisturbed by continental heating. Apparently this subsidence and horizontal divergence take place most effectively in the lower layers, since the lapse rates are relatively steep aloft. However, it is possible that it has subsided just as much at high levels, but that the lapse rate before subsidence set in was very steep.

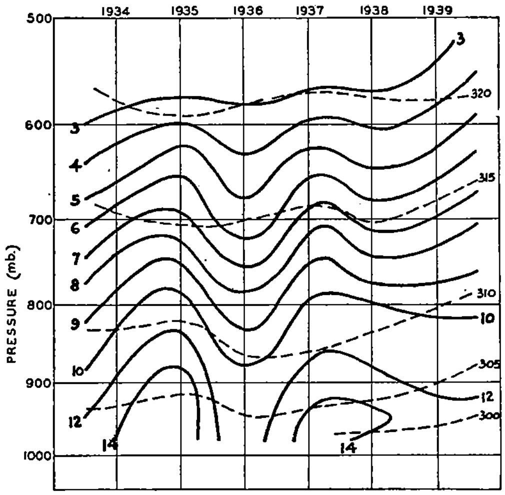

FIG. 40. Mean values of mixing ratio (solid lines) and potential temperature (broken lines) over Oklahoma City, Oklahoma, for the months of August 1934–1939.

While the unicellular pattern appears especially well related to drought conditions in the mid-west, it is not the only drought producing pattern. For example, in July 1935, there were large deficits of precipitation in this area, and the isentropic chart showed two distinct eddies. But in all drought months studied thus far the western anticyclonic cell is well developed and there is a strong flow of dry subsiding polar air into it. The dry above moist stratification is thus maintained for long periods.

e. *The Rain-Producing Flow Patterns.* The mean isentropic flow patterns associated with above normal summer rains in the mid-west stand out in sharp contrast to the drought-producing patterns. Whereas a well-developed anticyclonic cell characterizes the latter type, in the former the western cell is generally very weakly developed or is completely lacking. In fact, it appears that in the rainiest periods the western cell is actually supplanted by a *cyclonic* circulation. A typical example of such a month is June, 1935, a cold, wet month in the mid-west. The isentropic flow pattern for this month (Fig. 41) shows that the moist tongue is completely dominated by the cold and dry cyclonic circulation to its north. Apparently in this month there was appreciable upslope motion (frontal activity) of this moist tongue along the isentropic surfaces. The configuration of the moisture lines with respect to the contour lines indicates this.

At the surface the conditions during such wet months are marked by a series of waves which develop along the quasi-stationary polar front, and the precipitation, of course, falls mainly on the cold side of the front

From the charts studied thus far there appears to be little effect of the stage of development of the western cell on the eastern cell, except in the case when the western cell becomes exceptionally well established (as in August, 1936 and 1938) when it spreads out so that the eastern cell appears to be displaced far to the east.

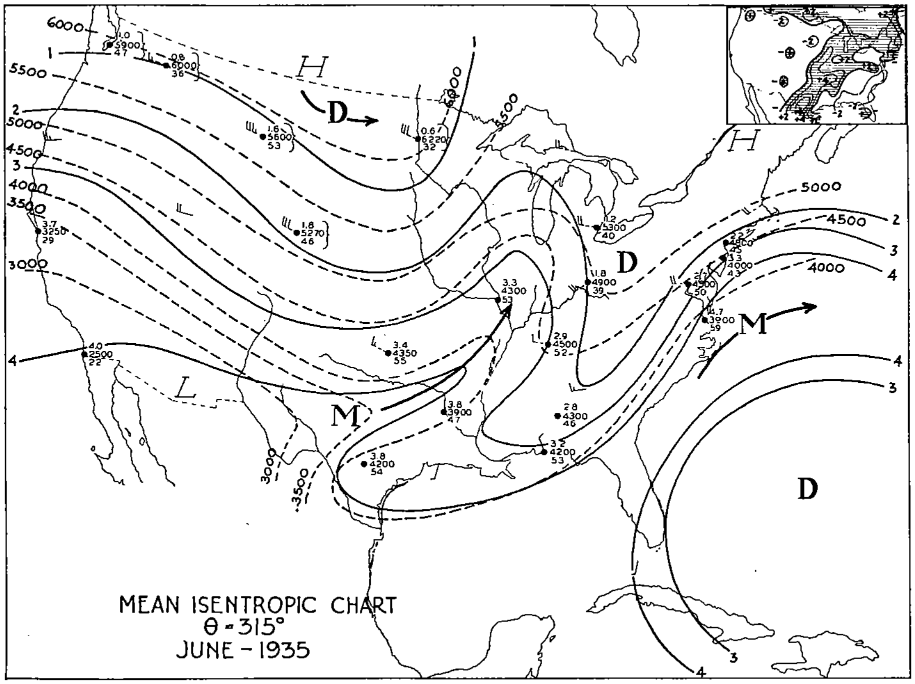

FIG. 41. Mean isentropic chart for June, 1935. (Inset—departure of precipitation from normal, expressed in inches.)

In the shear zone between the dry cool air of the eastern edge of the western anticyclonic cell and the moist current on the western edge of the eastern eddy there are often found heavy amounts of precipitation, for here the upslope effect is superimposed upon the normal shower condition of the moist tongues. A good example of this excess precipitation is afforded by the chart for August, 1935 (Fig. 42), where excesses of precipitation of as much as 10 inches were recorded in northern Florida.

f. Forecasting Factors for the General Summer Weather of the Mid-West. In searching for indices which might foreshadow the general nature of the summer temperature and rainfall of the mid-west, a careful study was made of the mean isentropic charts for the summer months and also for the month of May. This study, however, was handicapped by the fact that upper air data upon which reliable mean charts may be constructed are available only since 1934. In spite of this lack of material with which to work, many interesting and promising possibilities were suggested by these charts.

First, since the mid-west summer weather is closely associated with the development of the western anticyclonic cell, the forecast may depend on recognizing what conditions encourage the formation of such a cell, and what conditions may indicate its development in advance.

The temperature and precipitation departures of the mid-west would be expected to be in close relation, in accordance with the foregoing discussion of the upper level anticyclonic cells. Thus the correlation between the total summer (June, July, and August) rainfall amounts in the mid-west (defined by the states Iowa, South Dakota, Nebraska, Kansas and Oklahoma) and the temperature over this area for the years 1892 to 1937 was −0.79. For this reason a method of forecasting the summer temperature departures of the mid-west would go a long way towards solving the problem of precipitation forecasts for this area, or vice versa.

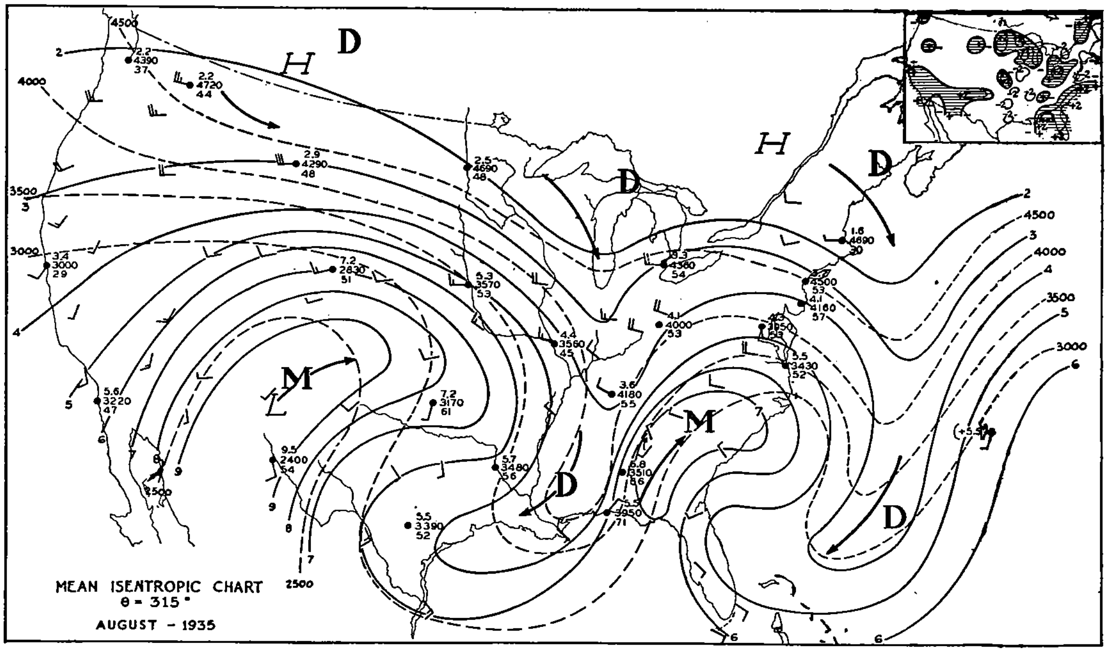

FIG. 42. Mean isentropic chart for August, 1935. (Inset—departure of precipitation from normal, expressed in inches.)

Through the study of the mean isentropic charts it was noted that during the period 1934–39 the western cell did not become well established until July of the years 1935, 1937 and 1939, while it was already well established in May, 1934, and fairly well established in May, 1936. These two latter years were the especially hot and dry summers of the mid-west. Table VI shows the stage of development of the western cell for various months of the period 1934–39.

It is readily seen from this table that there is some degree of persistence to the anticyclonic cell once it is established. After the cell becomes well developed, it appears to stay well developed (or fairly well developed) for a period of a few months. Thus, in 1934, the cell, established in May, persisted through August. In 1938 it was not well established until August, but once established, it persisted through October. The persistence of these upper level cells from week to week and from month to month implies a stability which seems to elude satisfactory explanation. As a possibility, we might consider the thermal effects once the cell is established. When well-developed, the isentropic surface over the cells assumes the shape of bowls whose lowest points are located over the mid-west. To the north, south, east and west the isentropic surfaces slope gradually upward. Now the anticyclone consists chiefly of dry T_s air, and this air appears to be fed sporadically into the cell from the north. Owing to the distribution of solenoids around it, there must be some "drive" existing on all sides, although chiefly to the north. Moreover, as long as the dry air predominates in the upper levels of the cell, moisture currents attempting to penetrate it laterally or from below will be suppressed or robbed of their moisture by lateral mixing. Besides, when the anticyclone is especially well-established, the circular rotation may prevent entry of new supplies of moist air from its jet source (over the southern Rockies). The thermal effect, then, may act undisturbed for long periods.

TABLE VI

Stage of Development of Western Cell 1934–39

	M	J	J	A	S
1934	G	G	G	G	P
1935	P	P	G	F	F
1936	F	G	G	G	P
1937	P	P	G	G	P
1938	P	P	F	G	G
1939	P	P	G	F	G

G—well developed
F—fairly well developed
P—Poorly developed

Since the May and June isentropic charts of 1934 and 1936 indicated well-formed cells and since these summers were especially noteworthy for severe mid-west droughts, the attempt was made to find surface indices in the May months which might indicate the state of development of the upper level cell. One possible index considered was the temperature in the region of the Great Lakes. Assuming that a cold winter and spring might result in a cold summer lake temperature, and that this would help stimulate a northwest flow aloft over the mid-west, a correlation was made of the accumulated departure from normal of the December through May temperatures of the area of the Great Lakes (determined by inspection of the monthly temperature departure charts of the Monthly Weather Review) with the following mid-west summer temperature as defined by the state averages for Iowa, South Dakota, Nebraska, Kansas and Oklahoma for June, July and August. For the period 1915 to 1930 the coefficient was 0.09. Using the April and May Lake Region departures, however, the coefficient rose to 0.33, and for the May Lake Region departures it rose to 0.55 for the same period. Further study indicated that well-developed western anticyclonic cells of the mean isentropic charts were generally associated with high temperatures in the mid-west, especially large positive departures from normal apparently centered around South Dakota. The departures from normal of South Dakota's May temperature were then used as in index of the stage of development of the western cell, which was correlated with the following summer mid-west temperatures (as defined above). For the period 1915–1937 this correlation coefficient was 0.60. Assuming that both the Lake Region departures and the South Dakota departures affected the following summer mid-west temperatures, a correlation was made between the sum of both May temperature departures and the following summer's temperature departure of the mid-west. This correlation was 0.70 for the period.

In order to make the data more readily available, a substitute had to be found for the May Lake Region departure of temperature for previous years. The Michigan May temperature departure was used, since it showed a correlation of 0.64 with the mid-west summer temperature, while the estimated Lake Region temperature departures showed +.55 correlation for the same period (1915–37). For the period 1892–1937 the correlations were:

Summer Mid-west temperature with South Dakota's May temperature 0.48; with Michigan's May temperature 0.37; with the sum of Michigan's and South Dakota's May temperature departures 0.46 (1892–1939).

It has been well known that there is appreciable persistence of temperature in the mid-west region from month to month in summer. The correlation of Iowa's May with the following June temperature for the period 1887–1937 is, for example, 0.31, and even higher month to month coefficients have been recorded for other states. This persistence has been noted by Charles D. Reed, who made use of the Iowa May and June temperature departures to forecast the general temperature departure for July and August. It should be noted, however, that the coefficients given above refer to a large area and are in the nature of "forecasting" coefficients for the average temperature of the entire summer.

It also was attempted to find other areas which might show sizable coefficients. Colorado's May temperature and the following summer's mid-west temperature were correlated at 0.46 (1892–1937) and Missouri's May temperature with the following summer's mid-west temperature at 0.24.

In view of these positive coefficients, it was believed that a multiple correlation between summer mid-west temperature and the May temperature for Michigan, South Dakota, Colorado and Missouri might be of value. This multiple coefficient for the period 1892–1939 was 0.53, the β coefficients being

Michigan	0.33
South Dakota	0.39
Colorado	0.18
Missouri	−0.38

Finally, since the May temperature distribution west of the Mississippi might influence the development of the western cell and hence the summer mid-west temperature, the product of each state (west of the Mississippi) with its May temperature departure from normal was found, and the sum total of these products was obtained. This final value was then correlated with the following summer's temperature departure in the mid-west and found to be 0.54 (period 1915–1939).

SECTION III. VERIFICATION OF FORECASTS

A. INTRODUCTION

The principal object of the Bankhead-Jones research as carried out at the Massachusetts Institute of Technology has been to develop practical methods of long-range forecasting by the process of deriving a theoretical or synoptic basis for forecasting some element, then testing its applicability with data for past years. Unless data is available for a long period of time, however, it is difficult to test the method on data which is entirely independent of that used in deriving the method. For this reason, and because climatological trends may alter the relations after they are once found from past data, the safest test of a forecasting method is always a test made with data occurring subsequent to the development of the method.

Since November, 1938, 5-day forecasts have been made each week in connection with the Bankhead-Jones project. Eight forecasters have participated, making individual forecasts. Some of these have made forecasts only during the school year, omitting the months June through September, and only four of the forecasters are the same for both winters. They will be referred to by number in this report—i.e., forecaster No. 1, forecaster No. 2, etc. It is not intended that the verification mark for these forecasts shall be used as an indication of the collective forecast value of the various ideas expressed in the preceding sections of the report. In most cases the methods have been tried individually with data for the past two years, and the results have been given along with the discussion of the method. Furthermore, some of the forecasters have consciously attempted in making their weekly forecast, to make strict application of one or more methods under trial, whereas others have used all available knowledge and forecasting experience to make the best possible forecast. Nevertheless, the slight gain in accuracy of the forecasts for the second year over that for the first year may indicate the advantage of having available during the second year methods and factors that were not yet formulated during the first year.

B. DESCRIPTION OF THE FORECASTS

Forecasts were made on Friday of each week of the sign of departure from normal of the mean temperature and precipitation for the 5-day period beginning at 7:30 A.M. Saturday for each of three sections. Each section was represented by the average of three stations; for the eastern section, Albany, Binghamton, and New York; for the southern section, Atlanta, Birmingham and Chattanooga; and for the western section, Kansas City, Springfield, Mo., and Wichita. A forecast was made also of the sign of the change of the element from the week just past to the coming week. Thus each complete forecast contained 12 items; three temperature departure forecasts, three temperature change forecasts, and the same for precipitation. On February 24, 1939, there were begun also forecasts of the numerical values of the 5-day mean temperature and 5-day total precipitation for the three sections. Once the numerical value is forecast, the sign of departure and change automatically follow, so that the two forecasts are not independent, and the verification mark of the numerical values probably should be given more weight in an interpretation of results.

Each week one of the forecasts, made by forecaster 4, was based entirely on a statistical method, the temperature forecast being based on persistence and the precipitation forecast based on the most probable value determined from a precipitation frequency distribution. The 5-day mean temperature departure was assumed to be a linear function of the mean temperature departure for the day preceding the forecast period. From past data a regression was calculated for each section of the form $y=bx$, where y is the 5-day mean temperature departure, and x is the mean departure for the day preceding the 5-day period. In use, the mean of the maximum temperature for 24 hours ending and the minimum temperature for 12 hours ending at 7:30 A.M. Friday was used for x. This equation thus gives a numerical forecast for temperature departure, which determines the sign of departure and sign of change to be forecast.

Frequency distributions of 5-day precipitation totals for the three sections show that for the eastern section, in the winter, the precipitation is below normal about 58% of the time. In the southern section it is below normal 63%, and in the western section 66% of the time. The serial correlation in 5-day precipitation is not significantly different from zero, hence a forecast of below normal seems to have a better chance of being correct than a forecast based on persistence. For that reason all precipitation departure forecasts made by forecaster 4 were negative (except for a few weeks at the beginning of the period). Precipitation change forecasts were based on another characteristic of the precipitation frequency distribution. The most probable amount of 5-day precipitation is an amount somewhat less than the normal, so that the change forecast should be always a forecast of change toward the most probable value, not toward the normal value. Weekly precipitation totals were available for 30 years for the three sections. From these data (divided by seasons) the sign of the change from week to week was tabulated as a function of the amount of precipitation during the first week, and a value determined such that for weekly totals below that value, the change to the next week was more often positive than negative, and for weekly totals above that value, the change was more often negative than positive. This value was multiplied by 5/7 and used in forecasting the change in 5-day precipitation during the second year. During the first year it was always forecast that precipitation would be nearer normal the next week.

C. Methods of Verification

The verification of forecasts of sign and of numerical values have been made separately. In verifying the forecasts of sign, they were separated into departure and change, and into temperature and precipitation. The percentage correct is given for each case, and wherever feasible the percentage correct that would be expected by chance over a long interval of time.

In the case of temperature departure forecasts, it has been assumed that positive and negative departures are equally likely, hence by chance 50% should be correct if many forecasts are made. However, for a small number of random forecasts, the observed percentage correct will have a distribution about 50% as a mean, and if the sample number is as great as 40, the distribution will be approximately normal, with standard deviation equal to $1/2\sqrt{n}$. On this basis, probabilities have been computed that as high a verification percentage as that obtained by the forecaster could have been obtained by chance. It was not possible to apply this method to any other group of forecasts, because the probabilities are not the same for all forecasts within a group. The method is questionable, even for temperature departure forecasts, because it does not take into account the effect of persistence. The probabilities obtained would tend to be too low on this account.

For forecasts of temperature change, the probability of a success depends on the magnitude of the departure the preceding week, since obviously if the departure is extremely large, a forecast of change toward normal has a very good chance of being correct. The same is true of precipitation change forecasts, except that the most frequent change is toward a value somewhat less than normal. No attempt was made then, to compute expected percentages for change forecasts, but it may be noted that the expected values would be greater than 50%.

For precipitation departure forecasts, the probability of a success depends on whether the forecast sign is positive or negative, and also on the region for which the forecast is made. For purposes of computing the expected percentage correct, a tabulation was made for each forecaster of the number of forecasts of each sign for each region. These were multiplied by the appropriate factor from Table VII, to obtain the expected number correct by chance, then the expected numbers were summed and divided by the total number of forecasts.

TABLE VII

PROPORTION OF POSITIVE AND NEGATIVE DEPARTURES IN 5-DAY TOTAL PRECIPITATION FOR THE THREE FORECAST SECTIONS, BASED ON DECEMBER, JANUARY AND FEBRUARY OF THE YEARS 1915 TO 1937

	Positive Departure	Negative Departure
East	.42	.58
South	.37	.63
West	.34	.66

For temperature and precipitation departures combined an expected percentage was obtained by averaging the separate values. This is correct when the number of forecasts is the same for both.

In addition to a comparison of actual and expected percentage verification, it is important in the case of precipitation forecasts to know how well the positive departures were forecast. As is shown by the statistical forecasts, a relatively high percentage can be obtained by forecasting negative departure exclusively. However the practical value of the forecasts obviously is increased enormously by the ability to forecast large amounts of precipitation, provided also that the forecaster does not give an excessive number of plus forecasts in order to be more sure of hitting the few positive departures. The proportion of correct plus forecasts was computed, also the percentage correct that would have been obtained if all forecasts had been negative. A series of good precipitation departure forecasts then must have a high percentage verification on the plus forecasts to indicate that few "false alarms" were given; the percentage verification of all forecasts must be higher than the expected percentage, and the total mark should be higher than the percentage that would have been obtained if all forecasts had been negative.

The principal test of the value of the numerical forecasts has been the correlation of forecast and observed departures from normal. Such a correlation, when squared, indicates the variation in observed departures which may be accounted for by the forecasts, but it does not completely measure the actual correspondence of forecast and observed departures. The standard deviation of the forecast departures may be small compared to the standard deviation of the observed departures, so that only a small departure is forecast when very large departures occur, and yet the correlation in such a case may be quite high. Along with the correlations there are given the standard deviations of both forecast and observed departures, and the ratio of the standard deviation of forecast to the standard deviation of observed departures. A group of forecasts can be considered as having more practical value if this ratio and the correlation are both high, than if only one is high.

Correlations were computed for each region separately, for both temperature and precipitation. A combined coefficient for all regions was obtained by converting the coefficients into Fisher's z, averaging the z's, then reconverting into a correlation coefficient. This averaging is not legitimate, however, if the separate coefficients are significantly different. The same restriction applies to the combined standard deviation for all three sections which was obtained by adding the sums of squares for the three sections, each corrected for the mean of the section, dividing by the total number of items, then taking the square root. For the precipitation forecasts, the standard deviations given have little meaning as such, due to the extremely skew distribution of precipitation values. Their usefulness here is simply as indicators of the relative "spread" of forecast and observed departure.

It will be of interest to know, for each forecaster, how the actual forecast errors compare with the errors that would have been obtained by forecasting the normal value each time. To obtain this comparison, the square root of the mean squared error (forecast minus observed) for the actual forecasts, and the root mean square difference between observed and normal, not corrected for the mean, were computed for each group of forecasts. In order for the forecasts to be better than forecasts of the normal, the ratio of the actual error to the normal error should be less than one.

D. RESULTS OF VERIFICATION

The results of the verification of the forecasts of sign are given in Table VIII. For each forecaster and each element, the table shows the number of forecasts falling in each of the categories

A)	Forecast sign plus	Occurred sign plus
B)	" " plus	" " minus
C)	" " minus	" " plus
D)	" " minus	" " minus

These are arranged as follows:

	Forecast		
Occurred	+	–	Total
+	A	C	A+C
–	B	D	B+D
Total	A+B	C+D	A+B+C+D

The sum A+C is the total number of plus occurrences, B+D is the number of minus occurrences, A+B is the number of plus forecasts and C+D is the number of minus forecasts. Likewise, the sum A+D is the total number of correct forecasts. The periods included in the two years are November 26, 1938, to May 31, 1939, for the first year and October 7, 1939, to April 10, 1940, for the second year. It will be noted that only forecasters 1, 3, 4 and 7 made forecasts for both years, of which No. 4 was the statistical forecast.

TABLE VIII

SUMMARY OF FORECASTS OF SIGN OF DEPARTURE AND CHANGE, SHOWING NUMBER OF FORECASTS CORRECT, NUMBER WRONG, PERCENTAGE EXPECTED CORRECT AND ACTUAL PERCENTAGE CORRECT

FORECASTER NUMBER	YEAR	TEMPERATURE DEPART.							TEMP. CHANGE					ALL TEMP.	PRECIPITATION DEPARTURE								PRECIP. CHANGE					ALL PRECIP.	PRECIP.&TEMP. DEPARTURES		ALL FORECASTS
		OCCUR'D	FORECAST +	FORECAST −	FORECAST TOT.	% CORR.	EXP. NO. CORR.	PROB.	OCCUR'D	FORECAST +	FORECAST −	FORECAST TOT.	% CORR.	% CORRECT	OCCUR'D	FORECAST +	FORECAST −	FORECAST TOT.	% CORRECT PLUS FORE.	% CORRECT ALL FORE.	EXP. % CORRECT	% CORRECT IF ALL MINUS	OCCUR'D	FORECAST +	FORECAST −	FORECAST TOT.	% CORR.	% CORRECT	ACTUAL % CORRECT	EXPECTED % CORRECT	% CORRECT
1	1938-39	+	28	4	32	64.6	24	.04	+	20	6	26	72.9	68.8	+	12	8	20	60.0	66.7	52.9	58.3	+	19	5	24	66.0	66.3	65.6	51.4	67.5
		−	13	3	16				−	7	15	22			−	8	20	28					−	11	12	23					
		TOTAL	41	7	48				TOTAL	27	21	48			TOTAL	20	28	48					TOTAL	30	17	47					
	1939-40	+	11	14	25	47.6	31.5	.70	+	26	8	34	76.2	61.9	+	6	14	20	30.0	55.6	54.7	68.2	+	26	1	27	75.0	65.0	51.6	52.4	63.5
		−	19	19	38				−	7	22	29			−	14	29	43					−	14	19	33					
		TOTAL	30	33	63				TOTAL	33	30	63			TOTAL	20	43	63					TOTAL	40	20	60					
2	1938-39	+	33	11	44	63.5	31.5	.03	+	27	10	37	61.9	62.7	+	10	17	27	43.5	52.4	53.0	57.1	+	24	14	38	66.1	59.2	57.9	51.5	61.0
		−	12	7	19				−	14	12	26			−	13	23	36					−	7	17	24					
		TOTAL	45	18	63				TOTAL	41	22	63			TOTAL	23	40	63					TOTAL	31	31	62					
3	1938-39	+	30	16	46	62.1	33	.05	+	26	11	37	71.2	66.7	+	13	14	27	40.6	50.0	50.1	59.1	+	24	10	34	66.2	58.0	56.1	50.0	62.4
		−	9	11	20				−	8	21	29			−	19	20	39					−	12	19	31					
		TOTAL	39	27	66				TOTAL	34	32	66			TOTAL	32	34	66					TOTAL	36	29	65					
	1939-40	+	18	13	31	58.7	37.5	.13	+	25	12	37	70.7	64.7	+	13	10	23	37.1	57.3	51.6	69.3	+	29	3	32	76.8	66.7	58.0	50.8	65.6
		−	18	26	44				−	10	28	38			−	22	30	52					−	13	24	37					
		TOTAL	36	39	75				TOTAL	35	40	75			TOTAL	35	40	75					TOTAL	42	27	69					
4	1938-39	+	32	26	58	55.6	40.5	.32	+	29	16	45	75.3	65.4	+	7	22	29	63.6	67.9	58.8	64.2	+	31	9	40	65.0	66.5	61.7	54.4	65.9
		−	10	13	23				−	4	32	36			−	4	48	52					−	19	21	40					
		TOTAL	42	39	81				TOTAL	33	48	81			TOTAL	11	70	81					TOTAL	50	30	80					
	1939-40	+	23	10	33	70.4	40.5	.0002	+	32	9	41	72.8	71.6	+	0	24	24	0.0	70.4	62.1	70.4	+	34	2	36	76.0	73.1	70.4	56.0	72.3
		−	14	34	48				−	13	27	40			−	0	57	57					−	16	23	39					
		TOTAL	37	44	81				TOTAL	45	36	81			TOTAL	0	81	81					TOTAL	50	25	75					
5	1939-40	+	23	10	33	76.5	40.5	.00001	+	32	9	41	84.0	80.2	+	3	21	24	33.3	66.7	59.8	70.4	+	32	4	36	74.7	70.5	71.6	54.9	75.5
		−	9	39	48				−	4	36	40			−	6	51	57					−	15	24	39					
		TOTAL	32	49	81				TOTAL	36	45	81			TOTAL	9	72	81					TOTAL	47	28	75					
6	1938-39	+	26	15	41	53.3	30	.60	+	24	11	35	71.7	62.5	+	19	7	26	54.3	61.7	48.2	56.7	+	23	8	31	67.8	64.7	57.5	49.1	63.6
		−	13	6	19				−	6	19	25			−	16	18	34					−	11	17	28					
		TOTAL	39	21	60				TOTAL	30	30	60			TOTAL	35	25	60					TOTAL	34	25	59					
7	1938-39	+	33	13	46	63.6	33	.03	+	26	11	37	72.7	68.2	+	11	16	27	44.0	54.5	52.9	59.1	+	27	7	34	73.8	64.1	59.1	51.4	66.2
		−	11	9	20				−	7	22	29			−	14	25	39					−	10	21	31					
		TOTAL	44	22	66				TOTAL	33	33	66			TOTAL	25	41	66					TOTAL	37	28	65					
	1939-40	+	22	8	30	66.7	33	.007	+	26	10	36	74.2	70.5	+	14	6	20	51.8	71.2	52.6	69.7	+	29	1	30	75.4	73.2	68.9	51.3	71.8
		−	14	22	36				−	7	23	30			−	13	33	46					−	14	17	31					
		TOTAL	36	30	66				TOTAL	33	33	66			TOTAL	27	39	66					TOTAL	43	18	61					
8	1939-40	+	17	7	24	66.7	31.5	.008	+	26	6	32	77.8	72.2	+	6	11	17	24.0	52.4	52.6	73.0	+	21	4	25	65.5	58.7	59.5	51.3	65.6
		−	14	25	39				−	8	23	31			−	19	27	46					−	16	17	33					
		TOTAL	31	32	63				TOTAL	34	29	63			TOTAL	25	38	63					TOTAL	37	21	58					

During the first year the temperatures in all sections were above normal much more frequently than they were below normal. This tendency is reflected in the temperature departure forecasts, which were predominately plus for all forecasters except No. 4. During the second year temperature departures were predominately negative, and this tendency again was recognized by all forecasters except No. 7.

In precipitation departure forecasts, however, some of the forecasters showed a tendency to forecast plus departure considerably more often than it occurred, particularly No. 3 the second year, and No. 6. On the other hand, No. 5 made too few forecasts of plus departure. The statistical method, as explained before, made no plus forecasts.

Assuming the 5% level of significance for the temperature departure forecasts, it is noted that of the four who made forecasts both years, only one, No. 7, had a mark significantly better than chance for both years. Nos. 4 and 7 showed an improvement for the second year over the first, whereas Nos. 1 and 3 obtained lower marks the second year. By far the best mark for either year was made by No. 5. Out of the twelve forecaster-years, 7 scores were significantly better than chance and only one, No. 1 the second year, obtained fewer correct forecasts than would be expected by chance. In the temperature change group, No. 5 again obtained the best mark, and it is noted that all percentages are considerably higher than in the temperature departure group, except No. 2. For all temperature forecasts combined, the highest mark the first year was obtained by No. 1, with No. 7 a close second. For the second year the highest mark was obtained by No. 5, with Nos. 4, 7 and 8 also having scores above 70%.

In the precipitation departure forecast group, three forecasters obtained marks lower than the expected value, No. 2, No. 3 the first year, and No. 8. Three other forecasters obtained marks more than 10% better than the expected percentage, No. 1 the first year, No. 6 and No. 7 the second year. The greatest improvement was shown by No. 7, whose score was only 1.6% above expected the first year and jumped to 18.6% above expected the second year. Four forecasters obtained marks better than would have been obtained if minus had been forecast all the time, No. 1 the first year, No. 4 the first year, No.6 and No. 7 the second year. The presence of the statistical forecast in this group is due to the 11 plus forecasts made at the beginning of the first year, of which a larger proportion was correct than would normally be the case. This also accounts for the presence of No. 4 the first year among those having more than 50% of the plus precipitation departure forecasts correct, the others being No. 1 the first year, No. 6, and No. 7 the second year. It should be noted that of these four, No. 1 made the same number of plus forecasts as occurred, No. 4 made only a few plus forecasts and No. 6 and No. 7 made one-third more plus forecasts than occurred. For the first year, the best precipitation departure forecasts, considering the three criteria of a good forecast as discussed on page 84, seem to be those of No. 1, while for the second year, the best forecasts are those of No. 7.

For all precipitation forecasts the highest percentage score the first year was obtained by No. 4, with No. 1 having practically the same score but having made a greater number of plus forecasts, hence more useful forecasts. For the second year No. 7 obtained the highest percentage score, and also made the most useful series of forecasts according to the other criteria.

The scores for temperature and precipitation departure combined show that No. 7 has the greatest margin of actual percentage correct over expected percentage correct, with No. 5 a close second. On all forecasts combined, No. 5 has the highest mark, with the statistical forecast second, and No. 7 a close third. From the point of view of practical usefulness, however, the forecasts made by No. 4 and No. 5 suffer greatly in comparison with No. 7, on account of their failure to forecast positive precipitation departures.

The results of the verification of the numerical forecasts are tabulated in Tables IX–XI. In Table IX are shown the correlations between forecast and observed temperature departures from normal, and the standard deviations of forecast and observed departures. It is noted that in some cases, for instance forecaster No. 1 the second year, the standard deviation of the forecast departures is fairly large compared to that of the observed departures, yet the correlation is relatively small. In other cases, for instance No. 4 the second year, the correlation is 0.52, but the standard deviation of forecast departures is less than one-third that of observed departures. The highest correlation is that for No. 7 the second year, and in this case also, the product of the correlation and the ratio of standard deviations, which is approximately equal to the coefficient of regression of forecast on observed, has its highest value. For forecaster No. 7, the forecast and observed departures have been plotted in Fig. 43. The diagonal line indicates the location of perfect forecasts. For points above the diagonal line, the forecast was algebraically

TABLE IX

THE CORRELATION BETWEEN FORECAST TEMPERATURE DEPARTURE FROM NORMAL AND THE OBSERVED TEMPERATURE DEPARTURE FROM NORMAL, ALSO THE STANDARD DEVIATION OF FORECAST AND OBSERVED DEPARTURES AND THE RATIO OF THE STANDARD DEVIATIONS

Temperature Departures from Normal

FORECASTER NUMBER	CORRELATION					ALL SECTIONS				
	Year	n	East	South	West	σ_F °F.	σ_0 °F.	$\frac{\sigma_F}{\sigma_0}$	r	$r\frac{\sigma_F}{\sigma_0}$
1	1938–39 1939–40	8 21	.17 .28	.58 .54	−.29 .19	2.56 4.41	5.38 7.60	.476 .580	.18 .35	.086 .203
2	1938–39	12	.35	.65	.23	4.20	5.66	.742	.43	.319
3	1938–39 1939–40	12 25	.20 .33	.14 .65	.09 .36	2.93 3.48	5.66 7.65	.518 .455	.14 .46	.073 .209
4	1938–39 1939–40	14 27	.24 .34	.23 .55	−.12 .63	1.61 2.38	5.73 7.68	.281 .310	.12 .52	.034 .161
5	1939–40	27	.46	.63	.56	3.78	7.68	.492	.55	.271
6	1938–39	11	.30	.51	−.18	1.88	5.58	.337	.23	.078
7	1938–39 1939–40	12 22	.42 .63	.67 .62	−.09 .41	2.77 4.85	5.66 7.17	.489 .676	.37 .56	.181 .379
8	1939–40	21	.43	.54	.27	5.02	7.62	.659	.42	.277

too low, and for points below the line, the forecast was too high. The one exceptionally large negative departure for the second year (South, −19°F) was forecast as a large negative departure, but of the two largest positive departures (both West, +18°F) one was forecast as positive and one was forecast as negative, both small in magnitude.

In Table X are given the correlations and standard deviations for precipitation departures. Here the highest combined correlation was obtained by No. 3 the second year, who also had the greatest improvement for the second year over the first year. However, the high coefficient is principally due to a high correlation for the western section, while the correlation for the southern section is slightly negative. Also the improvement in the forecasts as shown by the correlation was partially offset by the smaller ratio of standard deviations the second year. Forecaster No. 7 obtained almost as high a correlation as No. 3, with a higher ratio of standard deviations and with somewhat more uniform coefficients for the three sections. Of all the forecasters, only Nos. 3 and 7 during the second year made forecasts having any appreciable correlation with observed.

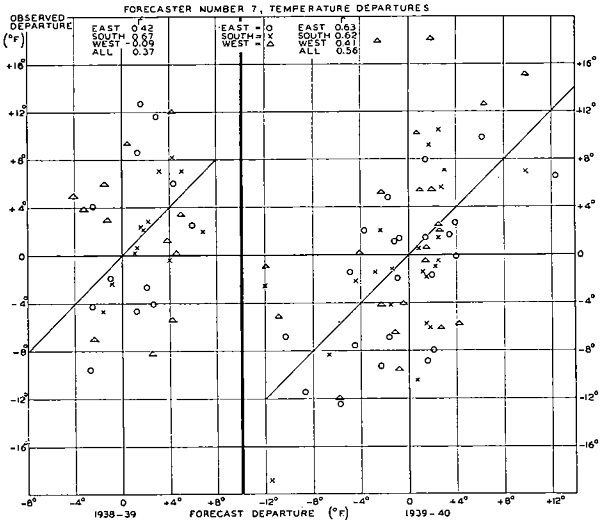

FIG. 43. Comparison of forecast and observed temperature departures for forecaster No. 7.

TABLE X

THE CORRELATION BETWEEN FORECAST PRECIPITATION DEPARTURE FROM NORMAL AND THE OBSERVED PRECIPITATION DEPARTURE FROM NORMAL, ALSO THE STANDARD DEVIATION OF FORECAST AND OBSERVED DEPARTURES AND THE RATIO OF THE STANDARD DEVIATIONS

Precipitation Departures from Normal

Forecaster Number	Correlation					All Sections				
	Year	n	East	South	West	σ_F in.	σ_0 in.	$\frac{\sigma_F}{\sigma_0}$	r	$r\frac{\sigma_F}{\sigma_0}$
1	1938–39	8	.12	−.11	.04	.20	.85	.23	.02	.01
	1939–40	21	.19	−.21	.10	.26	.57	.45	.03	.01
2	1938–39	12	.48	−.27	−.59	.20	.76	.27	−.14	−.04
3	1938–39	12	−.38	−.30	−.05	.49	.76	.65	−.25	−.16
	1939–40	25	.23	−.06	.74	.20	.55	.37	.36	.13
4	1938–39	14	.15	−.07	.07	.02	.74	.03	.05	.002
	1939–40	27	.17	.11	.15	.03	.55	.05	.14	.01
5	1939–40	27	.19	.12	.08	.14	.55	.25	.13	.03
6	1938–39	11	.11	.42	.22	.18	.77	.24	.26	.06
7	1938–39	12	.50	−.31	.23	.24	.76	.32	.15	.05
	1939–40	22	.35	.12	.51	.21	.56	.38	.34	.13
8	1939–40	21	.04	−.32	.19	.20	.50	.40	−.03	−.01

For No. 7, the forecast and observed departures of precipitation have been plotted in Fig. 44. The scatter about the line for perfect forecasts is considerably greater for precipitation than it was for temperature in Fig. 43. Of the six large positive departures (above one inch) during the first year, three were forecast with the correct sign, and of the five during the second year, again three were forecast with the correct sign.

Table XI shows the square root of the mean squared error of the actual forecasts, and of the error that would have been obtained by forecasting normal exclusively. For temperature, the ratio is less than one for everyone forecasting the second year, that is, the forecasts were better than normal forecasts. For the first year only forecasters 6 and 7 obtained ratios less than one. For precipitation only the forecasts of No. 6 were better than normal the first year, whereas the second year Nos. 4, 5 and 7 made forecasts better than normal.

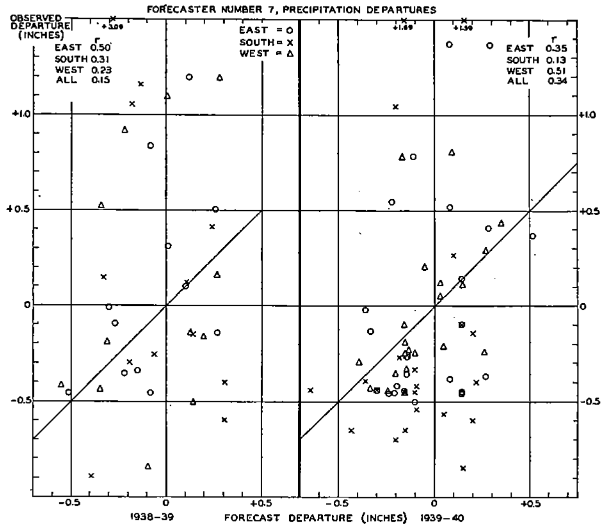

FIG. 44. Comparison of forecast and observed precipitation departures for forecaster No. 7.

Because of the difficulty of weighing the different methods of verification and of combining marks for the various forecast groups, it is hardly possible to say which of the forecasters obtained the highest composite score. It is sufficient to say that for temperature forecasts No. 5, No. 7 and No. 8 made forecasts which satisfied all the criteria of useful forecasts, and for precipitation, only No. 7 satisfied all criteria which have been discussed. However, this should not be taken to indicate that the other forecasts were not useful forecasts. For instance, the temperature forecasts of No. 2 fail only to satisfy the criterion that the numerical forecasts shall be better than normal forecasts. The statistical method has shown a remarkable ability to obtain high percentage verification on both temperature and precipitation forecasts of sign, but is not capable of making useful numerical forecasts, particularly of precipitation.

TABLE XI

SHOWING THE SQUARE ROOT OF THE MEAN SQUARED ERROR OF THE NUMERICAL FORECASTS, AND THE SQUARE ROOT OF THE MEAN SQUARED DIFFERENCE BETWEEN OBSERVED AND NORMAL, AND THE RATIO OF THE TWO

FORECASTER NUMBER	YEAR	TEMPERATURE			PRECIPITATION		
		Forecast Error	Observed Departure	Ratio	Forecast Error	Observed Departure	Ratio
1	1938–39	6.5	5.6	1.15	.90	.89	1.01
	1939–40	7.6	8.0	0.95	.60	.58	1.04
2	1938–39	6.0	6.0	1.01	.86	.78	1.07
3	1938–39	6.2	6.0	1.04	1.00	.78	1.28
	1939–40	7.1	8.0	0.88	.57	.56	1.03
4	1938–39	6.2	6.1	1.03	.76	.75	1.01
	1939–40	6.9	8.0	0.86	.54	.56	0.98
5	1939–40	6.5	8.0	0.82	.54	.56	0.98
6	1938–39	5.5	5.8	0.95	.77	.80	0.96
7	1938–39	5.7	6.0	0.96	.82	.78	1.05
	1939–40	6.4	7.5	0.86	.53	.56	0.94
8	1939–40	8.0	8.1	0.98	.58	.53	1.08

It is desired to draw some conclusion about the relative merit of the forecasts for the two years. For this purpose only those four forecasters who participated both years can be considered. For No. 1, whereas the percentage verification of signs in general was smaller the second year, the numerical forecasts were somewhat better, particularly for temperature. For No. 3 the percentage verification decreased on temperature but increased on precipitation, and the numerical forecasts were considerably better for both temperature and precipitation. The increase in the score of No. 4 on all points may be attributed principally to a fortuitous increase in the effect of persistence and to an increase in the proportion of negative precipitation departures for the second year. The verification score of No. 7 also increased in all particulars for the second year.

Thus two forecasters show increases in the percentage score for temperature forecasts of sign; three forecasters show increases in percentage score for precipitation; and all four forecasters obtained better verification scores the second year on the numerical forecasts. Hence, while it cannot be proved statistically that the increases are significantly better than chance, still it seems likely that in view of the new methods developed, these increases have been the result of real increases in forecasting skill.

CONCLUSION

From the scope of the above report it will be recognized at once that a wide range of more or less related topics in theoretical, synoptic, and statistical meteorology have received some consideration and investigation in connection with the five-day forecast project at M.I.T. It is quite impossible at the present time even to estimate the ultimate value of the work which has been carried on. However, some general conclusions and suggestions seem to be in order.

In the first place, it becomes evident at once on the perusal of this report that the results of much of the work, especially in the statistical-synoptic field, have been negative rather than positive, at least as far as practical applicability in forecasting is concerned. That result was anticipated in an investigation of this kind. Two general conclusions, both negative in import, which may be drawn from the past three years' efforts, are: (1) That most of the long range forecast methods (for periods extending more than half a week ahead) which have been applied up to the present, are not accurate enough to be of much practical value. The margin of accuracy which they possess above that which might be obtained by routine statistical forecasts is rather small. (2) That it appears that very little improvement of such forecasts can be made by synoptic or statistical methods except as these methods are based on improved physical and mathematical concepts of the dynamics and mechanics of the circulation of the earth's atmosphere.

A considerable number of specific synoptic or statistical relationships, more or less hypothetical in nature, were subjected to statistical checks. For the most part the results of such checks were discouraging from the point of view of utilization of the relationships in practical forecasting. There were, however, some notable exceptions, which have been incorporated into the M.I.T. five-day forecast practice. This applies especially to the use of aerological data in the North American area, and to certain relationships between the zonal index and the behavior of the centers of action.

There are three principal ways in which it is felt that definite positive results have been obtained from the project. In the first place, it has offered a strong stimulus to the development of a practical working concept or model of the general circulation. Real progress has been made in this most fundamental of all meteorological problems. This appears principally in the contributions of Professor Rossby which have been discussed in Section I of this report. Notable among these contributions are (1) his conception of the role of horizontal mixing in the general circulation and the consequent development of the technique of isentropic analysis; (2) his explanation of the movement of perturbations in the zonal westerlies as related to the behavior of the centers of action; and (3) his introduction of the concept of potential isentropic vorticity as a conservative quantity of practical significance in synoptic meteorology.

In the second place, positive results have been obtained from this project through the immediate practical application of these new working concepts of the general circulation to the practical problems of synoptic analysis and weather forecasting. In this way all new ideas are subjected to an immediate practical synoptic and statistical check on the one hand, while at the same time the development of the theory is to some extent guided by synoptic observation. This correlation of the two methods of attack on the forecast problem has doubtless been responsible to a large degree for the considerable percentage of statistically significant relationships which were found among those investigated and reported in Section II.

In the third place, the actual five-day forecast practice has been of great value to those partaking in it in building up a background of experience and in developing their appreciation of the synoptic possibilities of an extension of the time and space range of practical weather forecasting. This is gained largely through the critical use of northern hemisphere maps. The fact that the five-day forecasts have shown a definite upward trend in verification during the past two years substantiates the reality of this benefit.

If the five-day forecast project at M.I.T. is to be continued, there is no very radical departure from the procedure of the past two years which suggests itself. Effort should be expended on the same three general phases of the work, essentially as heretofore. That implies about the following program:

1. Continued emphasis on the study of the mechanics of the general circulation, by every available means, theoretical, synoptic, and statistical. The entire development of long range weather forecasting as a science depends on further progress in solving this problem.
2. A continuation of the synoptic and statistical checking of theories, principles, or hypotheses growing out of the study of the general circulation. There is still much to be done in the practical utilization of the ideas already developed. The weather on the Pacific coast seems to be more closely related to the state of the general circulation than has been realized until recently. Only a beginning has been made in the synoptic application of Rossby's recent vorticity concept. There is a great deal to be learned from an extension of the aerological charts, at least as soon as the re-establishment of normal world conditions makes the observations available. It appears also that there is much to be gained from a comparative study of past records for different years having widely varying characteristic features of the general circulation. This involves the question as to why some relationships which hold very closely on certain years almost vanish other years.
3. A continuation of the five-day forecasts, possibly even with an extension of the time range if developments warrant it. The preparation of successful long-range forecasts constitutes both the goal and the ultimate test of the practical value of the entire project. Consequently this part of the work has to be maintained to the fullest. It is probable that a verification system for the forecasts can be devised which will reflect more closely the true forecasting skill of the forecasters, as distinct from skill in capitalizing on the statistical probabilities.

There is little point in going into more detail now as to suggestions for future work under this project. The detailed program has always to be developed with the progress of the project. It should be evident, however, that much work remains to be done both in the further development of long range forecasting and in the complete utilization of the results of the work already completed.

APPENDIX

A list of the principal charts, diagrams, and circulation indices that have been prepared and computed in connection with the Bankhead-Jones research is given below, together with the dates covered by each type of chart. The list is under three headings, namely:

1) Charts and maps comprising the whole Northern Hemisphere.
2) Charts and maps comprising the continent of North America.
3) Circulation Indices.

1. Northern Hemisphere charts.
   - a. Northern Hemisphere daily weather maps
     November 15, 1936 to February 19, 1938
     October 15, 1938 to August 31, 1939
   - b. Northern Hemisphere 7-day mean pressure
     August 2, 1932 to April 25, 1933
     November 15, 1936 to October 15, 1938
   - c. Northern Hemisphere 5-day mean pressure
     October 15, 1938 to August 31, 1939
   - d. Northern Hemisphere seasonal mean pressure
     Spring season (March–May) 1921 to 1930 inc.
     Summer season (June–August) 1921 to 1930 inc.
     Autumn season (September–November) 1921 to 1930 inc.
     Winter season (December–February) 1921–22 to 1929–30 inc.
     } Computed at Washington
   - e. Northern Hemisphere 7-day mean pressure change
     January, 1937 to January, 1938 inc.
   - f. Northern Hemisphere 5-day mean pressure change
     October 15, 1938 to May 31, 1939
   - g. Northern Hemisphere zonal mean pressure profiles (daily)
     November 19, 1938 to May 31, 1939
   - h. Northern Hemisphere zonal mean pressure profiles (5-day mean)
     October 15, 1938 to August 30, 1939
   - i. Northern Hemisphere zonal mean pressure profiles (monthly mean)
     November, 1938 to May, 1939
   - j. Northern Hemisphere mean zonal velocity profiles (5-day means)
     October 15, 1938 to May 31, 1939

2. North American charts.
   - a. Surface pressure (60° west to 180° west) 7-day means
     October 15, 1939 to April 11, 1940
   - b. Surface pressure, change in 7-day means
     October 13, 1939 to April 11, 1940
   - c. 3-kilometer pressure, 5-day means
     October 22, 1938 to September 14, 1939
   - d. 3-kilometer pressure, 7-day means
     September 15, 1939 to April 11, 1940

e. Daily isentropic charts
   September, 1938 to April 11, 1940
f. 5-day mean isentropic charts
   June, 1938 to August, 1938
   October 22, 1938 to September 14, 1939
g. 7-day mean isentropic charts
   September 15, 1939 to April 11, 1940
h. Monthly mean isentropic charts
   January, 1935
   January, 1936 to April, 1936
   May to September, for all years 1934 to 1939
   October, 1938
   December, 1936
i. Seasonal mean isentropic charts
   Spring mean (March–May) 1936
   Summer mean (June–August) 1934 to 1939
   Autumn mean (September–November) 1936
   Winter mean (December–February) 1935–36, 1936–37
j. Resultant winds aloft
   Summer (June, July, August) normal 750 meters
   1500 meters
   3000 meters
   4000 meters
k. Weekly Precipitation Departure charts for United States
   Periods of maximum and minimum zonal index—October to March for years 1932–33, October, 1936 to March, 1939
   All weeks during June, July, August for 1934–1938
   All weeks October, 1939 to March, 1940

3. Circulation Indices.

a. 35°–55° Zonal Circulation Index (weekly means)
   August 2, 1932 to April 3, 1933
   June 5, 1934 to October 1, 1934 } Computed in part
   June 4, 1935 to October 18, 1938 } at Washington
   October 18, 1938 to May 29, 1939
b. 35°–55° Zonal Index (5-day means)
   October 15, 1938 to August 30, 1939
c. 45°–65° Zonal Index (5-day means)
   October 15, 1938 to May 27, 1939
d. Meridional Total Velocity Index (daily and 5-day mean)
   November 5, 1938 to May 31, 1939
e. Meridional Net Velocity (daily) for each 5° by 5° square from 20° to 75°N latitude, November 19, 1938 to March 31, 1939.