R B Montgomery

An Introduction to the Study of

Air Mass and Isentropic Analysis

By

JEROME NAMIAS

FIFTH REVISED AND ENLARGED EDITION

with

CONTRIBUTIONS

by

TOR BERGERON

BERNHARD HAURWITZ

GRAHAM MILLAR

ALBERT K. SHOWALTER

ROBERT G. STONE

AND

HURD C. WILLETT

Edited by

ROBERT G. STONE

October, 1940

E AMERICAN METEOROLOGICAL SOCIETY, MILTON, MASS.

Price $1.25, postpaid

Given in Loving Memory of

Raymond Braislin Montgomery

Scientist, R/V Atlantis maiden voyage

2 July - 26 August, 1931

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Physical Oceanographer

1940–1949

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1950–1960

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The Previous Editions of This Work

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"AN INTRODUCTION TO THE STUDY OF AIR MASS ANALYSIS"

First Edition . .	Sept., 1934-May, 1935
Second Edition . . .	September, 1935
Third Edition . . .	August, 1936
Fourth Edition . . .	October, 1938

TABLE OF CONTENTS

Corrigenda for "An Introduction to the Study of Air Mass and Isentropic Analysis," by J. Namias and others, 5th ed.

Page 5, table at bottom of page, lines under "Condition" and "Type of equilibrium", should be deleted and the following should be substituted (the preferred terminology is italicized:—

$\alpha < \beta$	(Absolutely) *stable*
$\alpha = \beta$	*Neutral for saturated air* (equal to wet-adiabatic lapse rate)
$\beta < \alpha < \gamma$	(Conditional:) *stable for dry air, unstable for saturated air.*
$\alpha = \gamma$	*Neutral for dry air* (equal to dry-adiabatic lapse rate)
$\gamma < \alpha <$ 3.42 C° per 100 m	(Absolutely) *unstable* (but not self-starting)
$\alpha =$ 3.42°/100 m	*Upper limit of mechanical stability*
[$\alpha >$ 3.42°100 m	*Mechanical instability*, or *auto-convective gradient* (self starting; hence this is not an equilibrium)]

Page 6, col. 2, line 8 of footnote, "J.-J. Jang" should read "Jaw."

Page 14. col. 1, line 11, "$T/P_1^{0.288}$" should read "$T/p_1^{0.288}$."

Page 14, col. 1, line 18, "wet-bulb temperature" should read "wet-bulb potential temperature."

Page 69, line 7, "Su" should read "Sc."

Page 71, top, line 4, delete "warm."

Page 150, col. 2, line 4, "stream function" should read "isentropic acceleration potential" (see BULLETIN A. M. S., Jan. 1941, p. 45).

Page 172, line 18 (in fine print) "frontal at" should read "frontal topography at."

Page 174, line 9, "15 km" should read "18 km or higher."

Page 174, line 10, "by 10 mb" should read "by 9 mb."

Page 175, line 1, "In Fig. 6, shown above is" should read "Fig. 6 above is". . .

Page 175, line 3, "tions shows" should read "tions. It shows."

Page 228, 2nd col. line 34, "to dry air" should read "to saturated air;" line 33, after "entire" insert "originally stable".

Page 229, 2nd col., lines 2 and 3 from bottom, "decreases" should read "increases."

Page 232, 2nd col., 3rd line from bottom delete "equal".

[*Supplement to the June 1941 Bulletin of the American Meteorological Society*]

Editor's Preface to 5th Edition

UNCEASING demand has again induced the Society to extend this convenient booklet into a 5th revised and enlarged edition. The text of the 4th edition is reprinted with numerous secondary annotations and changes to indicate some of the present attitudes or practices that depart from those stated or implied in the previous editions. The practices in other countries are so diverse that no special note of them could be taken, but extensive citations of foreign literature are given in the Bibliography.

At the time the 4th edition appeared (Oct. 1938) a new technique and point of view, known as isentropic analysis, was under promising experimental development and it was anticipated by the authors and editor that any future edition of this "Introduction" would have to take account of the new method. Already in 1939 isentropic analysis was so generally practiced in U. S. A. that plans were laid to add an introductory chapter on the technique. Mr. Namias was engaged by Prof. Sverre Petterssen to prepare such a chapter for his excellent book "Weather Analysis and Forecasting" recently published. We are able to offer a slightly modified form of this chapter in our 5th edition, through the kind permission of Prof. Petterssen and of the McGraw-Hill Book Co., Inc., of New York.

The Bibliography in the 4th edition has been so widely appreciated that an effort is made to improve it materially in this new edition. Besides bringing it down to date, a great many more entries are added and the whole arranged conveniently by subjects. Finally, additional illustrations are provided in the appendix.

This opportunity has been seized to correct a few typographical and other errors which unfortunately passed in the 4th edition; the editor wishes to thank the numerous individuals who have kindly called our attention to errors and offered suggestions for improvement. The basic part of this introduction remains rather elementary, but many more technical annotations are provided for the numerous students who have the background to enter a little deeper into the subject. However, we have not attempted to enlarge the work into a textbook of synoptic meteorology. It is assumed that the reader is familiar with the elements of meteorology and with the general conception of the weather map as given in numerous and readily available textbooks, to which this booklet is only an adjunct of certain newer topics not particularly well-treated outside a few technical and expensive works.

The material available for direct analysis of upper-air conditions has lately reached a new high. There are now 34 regular aerological stations in the U. S., mostly using radiosondes, the largest network of its kind in the world and in history. The *adequate* use of such data in *forecasting* is being tried here for the first time anywhere and will call for an increasingly quantitative and physical approach.

The editor wishes to acknowledge the very generous assistance in reading proof and valuable advice on many points given by Prof. Charles F. Brooks, Secretary of the Society, who has handled the arrangements for publishing this work. Thanks are due Miss Edna Scofield for aid in editing and reading proof.—*Robert G. Stone, Sept. 1940.*

From the Preface to the 4th Edition

When these sketches were first prepared it was thought that the subject would develop so rapidly that any attempt to simulate a well-rounded and comprehensive treatise, even for beginners, would not be justified nor feasible. Needless to say the continued warm response and wide influence which the work has had has left the authors and editors with a sense of responsibility for which they had not bargained. The authors feel that most of the early principles of air mass analysis still bear a fundamental value for synoptic practice, so that extensive revisions have not been necessary in this new edition, though references are made here and there to some of the points that have been particularly altered or questioned in light of more recent developments.

From the practical point of view, the beginner in America now has much better opportunities to "pick up" experience through his own efforts than when this "Introduction" first appeared over three years ago, when no competently analyzed air-mass weather maps were available outside of a few institutions and special services and none to the public. At present, thanks to the remarkable changes in U.S. Weather Bureau practice since 1934, such maps may be inspected by anyone at a large number of airport and city offices of the Bureau, and many of their personnel can now produce or interpret analyses acceptably, while at the Central Office in Washington competent and fundamental research is being carried on. (The problem of rapidly training a large organization in such a different technique is admittedly difficult).

The present "Introduction", however, should not be regarded as one to the whole field of synoptic meteorology. For such a serious ambition one should study physical meteorology as well, and there are excellent general texts such as Humphreys' "Physics of the Air", Brunt's "Physical and Dynamical Meteorology", Taylor's "Aeronautical Meteorology", "Byers' "Synoptical and Aeronautical Meteorology", [and now (1940) also Petterssen's "Weather Analysis and Forecasting", Sutcliffe's "Meteorology for Aviators", and "The Admiralty Weather Manual",] to lighten the road. But to that large audience which desires only a brief, authoritative, and inexpensive "first reader" in this fascinating concrete way of looking at the weather, this booklet is offered again in the hope that it will continue the instrument for wide dissemination of modern meteorological principles which it has been.—*Robert G. Stone,* Oct. 1938.

An Introduction to

The Study of Air Mass and Isentropic Analysis

By Jerome Namias

Introduction to the 5th Edition

The system of weather analysis developed chiefly by the Norwegian school of meteorologists and referred to as "Air Mass Analysis" in the United States, has received widespread adoption throughout the meteorological services of the world. In addition to the group of professional meteorologists who employ these methods as the foundation of their activities in synoptic meteorology, there has developed a large group of people whose interests are so intimately associated with meteorology that it is necessary for them to possess more than a fragmentary knowledge of the physical processes of the atmosphere. The increasing number of aviation enthusiasts is only one of these groups. Many such people, and indeed, many practicing professionals at present engaged in meteorology, have not had the time nor the opportunity to carry on an organized collegiate program of study in modern synoptic meteorology, and for this reason have felt the need for some simplified presentation of the fundamentals upon which the science rests. It has been the purpose of this series of articles to fulfill this gap in such a manner that these students may be able to obtain a *physical* picture of basic weather processes without first having to possess a mastery of advanced physics and mathematics. Since weather forecasting is still quite removed from the quantitative stage, and since a qualitative evaluation of the various entering factors constitutes a large share of the forecaster's technique, it is possible to develop a moderate degree of forecasting ability with an understanding of the physical processes as described in these articles.

The actual technique of air mass and isentropic analysis can hardly be imparted adequately by written material. It requires a more personalized guidance by an experienced analyst.

While textbooks in modern synoptic meteorology have been vastly improved since many of these articles were first written, notably by the works of Byers, Taylor and Sutcliffe, there has continued a demand for these articles, and more recently for some similar presentation dealing with the new method of upper-air analysis along the surfaces of constant entropy. Professor S. Petterssen (of M. I. T.) and the McGraw-Hill Company have been so kind as to permit me to publish here, with some small alterations, the chapter on isentropic analysis originally prepared for his textbook on *Weather Analysis and Forecasting* (McGraw-Hill, N. Y., 1940).

I. CONDITIONS OF ATMOSPHERIC STABILITY: LAPSE RATES

A. Vertical Displacement of a Particle

It has long been known that vertical motions in the atmosphere are of great significance in that practically all precipitation may be ascribed to the condensation brought about through expansional cooling of rising air. It can be shown that the amount of precipitation possible through the mixing of currents of air possessing different temperature and moisture characteristics is very small, and that the precipitation resulting from this cause must be negligible in comparison with other causes. Furthermore the theory of fronts and air masses is based upon atmospheric discontinuities, which are simply zones of rapid transition of the various meteorological elements. It is assumed that these zones of transition are comparatively free from large scale mixing, the individual large scale air currents flowing side by side or above one another without appreciable mutual drag.

Granting the importance of vertical motion in the atmosphere a discussion of the factors which tend to aid or hinder such motion is in order. This leads to the problem of stability. Here the term stability is used in its physical sense; if an air particle tends to remain in, or return to, its former position following a displacement, the condition is termed stable; if displacement results in a tendency to further movement of the particle from its original position, the original condition is designated as unstable; and finally, if the particle neither resists nor assists displacement, the condition is one of neutral equilibrium.

B. Types of Equilibrium

In the case of the atmosphere there are four principal types of equilibrium to be considered when we are concerned with the vertical displacement of a selected particle through a layer of the atmosphere having known characteristics. These types of equilibrium are:

1. Stable
   - a. With respect to dry air[1]
   - b. With respect to saturated air
2. Unstable
   - a. With respect to dry air
     - [1. Mechanical]
   - b. With respect to saturated air
3. Neutral
   - a. With respect to dry air
   - b. With respect to saturated air
4. Conditional

Another case, that of convective equilibrium, will be treated independently in a future article, since it concerns the displacement of layers of the atmosphere rather than the displacement of individual particles of air through a given layer.

It is obvious that if a particle of air is lifted it must expand against the decreasing pressure so that the pressure within and surrounding the particle must be equal; if it sinks it must be compressed. It is assumed that these changes take place without the transfer of heat either from the moving particle to its surroundings or vice versa. Such a thermally insulated process is termed adiabatic. If the particle expands it does work; if it is compressed work is done upon it. Thus there must be a conversion of mechanical energy into realized heat if the particle sinks, while heat must be converted into mechanical energy if the particle rises. By means of thermodynamics it can be shown that the relation between temperature and pressure in an adiabatic displacement of an unsaturated particle is as follows:

$$\frac{T_1}{T_2} = \left(\frac{p_1}{p_2}\right)^{0.288}$$

where T_1 is the original temperature of the particle at the pressure p_1, and T_2 is the temperature it assumes at the pressure p_2. This is Poisson's equation.

It is generally more convenient in aerological studies to refer to the adiabatic changes in temperature with respect to changes in elevation. From

[1]The term "dry air" in synoptic meteorology simply means unsaturated air.

Poisson's equation and the hydrostatic equation (expressing the relation between pressure, density, and height), it is possible to obtain the rate of cooling of a rising air particle owing to its change in elevation. The result is the convenient rate of 1 C deg. per 100 m. This rate is not strictly constant; it depends upon the amount of moisture within the unsaturated air particle as well as the temperature of the surrounding air through which it is displaced. However, these effects are relatively small and tend to counteract each other. Hence, for all practical purposes, they may be neglected, the adiabatic rate of change of temperature being taken as 1 C deg. per 100 m. change in elevation. This is commonly known as the dry adiabatic lapse rate, or dry adiabat. (Lapse rate is defined as the rate of change in temperature with respect to height. Unless preceded by the qualifying term "adiabatic," lapse rate refers to the *existing* difference of temperature per unit of height within a selected layer of the atmosphere.)

Thus far we have considered the vertical displacement of an unsaturated particle of air. Once the particle becomes saturated the latent heat of condensation must be taken into account, for it supplies heat to the rising mass and therefore lessens the rate of cooling due to expansion. The lessening of the adiabatic cooling effect depends upon the liberated heat of condensation, which in turn depends upon the amount of liquid water condensed. But as the particle

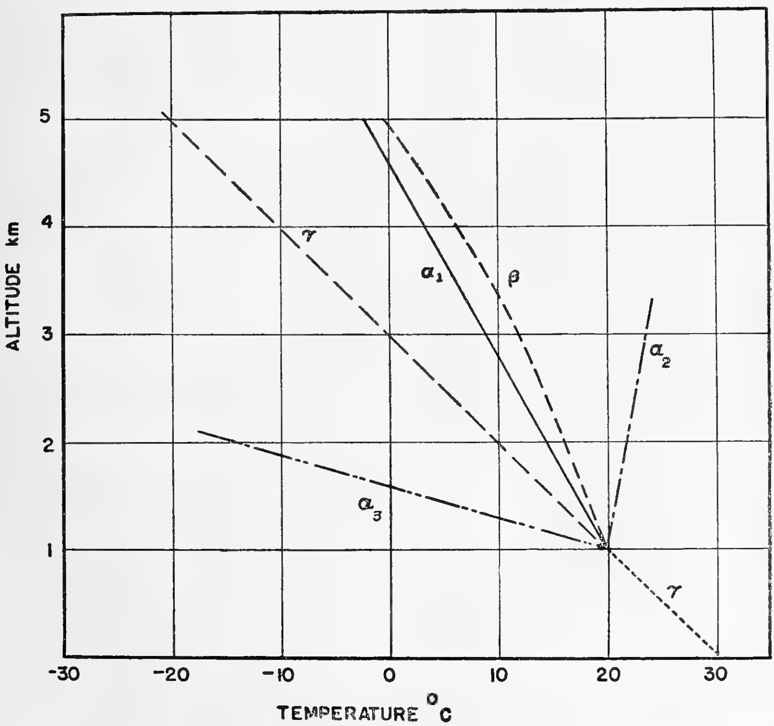

FIG. 1. TYPES OF LAPSE RATES

continues to rise, its temperature falls, so that the total quantity of water vapor possible within the volume of rising air becomes less and less. Therefore the rate of cooling of the saturated mass becomes greater and greater, until at high levels, where the moisture content of the rising air is almost negligible, its rate of adiabatic cooling is practically the same as that for dry air.

We are now prepared to deal with the types of equilibrium as outlined above.

1. *Stable.* The rate of cooling for an unsaturated particle of air rising through the surrounding atmosphere is given by the broken line γ in fig. 1. This is a straight line since the rate is 1 C deg. per 100 m. Lines drawn parallel to γ would represent other dry adiabats at different temperatures. If we assume the observed vertical temperature distribution (the lapse rate) above 1000 m to be represented by the line α_1, it is at once clear that a particle of air taken from any position on the line α_1 and brought up or down will immediately find itself of a different temperature and hence different density from its surroundings. It will consequently tend to return to its original position. For example let us take a particle at 1000 m. where the temperature is 20° C. If we bring this particle up to 2000 m., it will follow the line γ and at 2000 m. will assume the temperature 10° C. The surrounding air at 2000 m., however, has the temperature 14° C., or 4 deg. warmer than the rising particle. Under this condition the particle must return to its former position, coming to rest at 1000 m, where its temperature is the same as the surrounding air. In a similar fashion it is easily shown that downward motion of individual particles originally lying along the line α_1 are hindered, the tendency being always to make the particle return to its original position. It is clear then that the line α_1 represents a stable lapse rate. In other words if the lapse rate is less than the dry adiabat the layer is stable for unsaturated air.

The rate of adiabatic cooling for a rising saturated particle of air is represented in fig. 1 by the line β. Note that β is a curved line, since the rate of cooling is dependent upon the heat of condensation as well as upon the expansion. Also note that the curve β tends to straighten, gradually approaching the slope of γ at upper levels, where the moisture content becomes smaller and smaller. If we now assume a lapse rate of α_2 from 1 to 3 km. it is clear that a rising particle of saturated air will follow the line β, and will at every stage in its ascent be colder than its surroundings. Thus it will tend to remain at its original position and the layer between 1 and 3 km. will, by definition, be termed stable with respect to saturated air. A lapse rate less than the saturated adiabat may be termed absolutely stable since it is stable whether the rising air be dry or saturated.

2. *Unstable.* Let us assume that the lapse rate between 1 and 2 km. has the form α_3 as in fig. 1. A particle of air displaced upward from any position on the line α_3 would follow parallel to the dry adiabat γ and obviously would be warmer than the surrounding air, level for level. Therefore it would continue to rise. The layer possessing the lapse rate α_3 is then unstable with respect to dry air, and since the saturated adiabatic lapse rate is always less than the dry, it is clear that this condition is even more unstable for saturated air. The line α_3 has been constructed to represent a special case of instability in which the density remains constant

with elevation. The density, of course, is a function of the temperature and pressure, and is slightly affected by the moisture content. In the atmosphere the pressure decrease with elevation is such that it nearly always overbalances the increase in density caused by the usually observed drop in temperature with elevation. If the temperature falls off sufficiently rapidly with elevation, however, a state will be reached wherein the density of the air is constant with height. If the lapse rate exceeds the value 3.42 C deg. per 100 m there must be an increase in density with elevation—obviously a very unstable condition. This particular case has been given various names, the best one probably being mechanical instability. α_3 represents a state of mechanical instability. This condition is never observed in the upper atmosphere, since it is such an unstable state. It is, however, frequently observed immediately overlying flat regions which become greatly heated during the summer daytime hours.

In the case of instability with saturated air the lapse rate must be greater than the saturated adiabat. In fig. 1 the line α_1 represents such a lapse rate between 1 and 3 km. It should be noted that the layer above 3 km. is not unstable for saturated air, since the rate of change of the temperature along α_1 above 3 km. is less than that along β.

3. *Neutral equilibrium.* With dry air this state is reached when the lapse rate is equal to the dry adiabat. Under this condition the rising particle will possess the temperature of the surrounding air at every stage in its ascent. Thus it will neither assist nor resist displacement. If the rising air is saturated then the condition for neutral equilibrium requires that the lapse rate equal the saturation adiabat.

4. *Conditional equilibrium. It was* pointed out that the lapse rate given by the line α_1 is stable for rising air, while between 1 and 3 km. it is unstable for saturated air, because the lapse rate α_1 lies between the saturated and the dry adiabat. When this state obtains the layer is said to be in conditional equilibrium. The condition is simply that the layer is unstable if saturated, but stable if unsaturated. This lapse rate is frequently observed in aerological soundings, and has been found to be important in the development of thunderstorms and showers. It should be noted that the conditional instability in the case of fig. 1 extends through the layer between 1 and 3 km., and no higher. Beyond 3 km. the lapse rate α_1 does not lie between the dry adiabat and the saturated adiabat *for* the *temperatures* at these elevations. The rate of change of temperature along the line β (above 3 km.) is greater than along the line α_1.

A summary of the above conditions is presented in algebraic form below: where α represents the existing rate of change in temperature with elevation (the lapse rate); γ, the dry adiabatic lapse rate; the β, the saturated adiabatic lapse rate.

Condition	*Type of equilibrium*
$\alpha < \gamma$	Stable for dry air
$\alpha < \beta$	Stable for both dry and saturated air (absolute stability)
$\alpha > \gamma$	Unstable for dry air (absolute instability)
$\gamma < \alpha < 3.42°$ C per 100 m.	Unstable for saturated air
$\alpha > \beta$	Unstable for both dry and saturated air
$\alpha = \gamma$	Neutral for dry air
$\alpha = \beta$	Neutral for saturated air
$\beta < \alpha < \gamma$	Conditional
$\alpha = 3.42$ C deg. per 100 m.	Upper limit of mechanical stability (density remains constant with elevation)
$\alpha > 3.42°$ C per 100 m.	All air is mechanically unstable (density increases with height); "auto-convective gradient".

II. CONSERVATIVE PROPERTIES OF AIR MASSES

An *air mass* is defined as an extensive body of air which approximates horizontal homogeneity. The properties of the air mass which are considered in this homogeneity are mainly temperature and moisture. Thus over the earth's surface one may observe large currents of air within which the temperature and moisture content remain fairly constant at any given level. These large scale currents of air have their origin at a *source region*—a large area characterized by sameness of surface conditions and evenly distributed insolation. Thus the northern part of Canada in winter may be considered as a source region in that it is practically snow-covered, and the amount of insolation received is almost evenly distributed over the entire area. A large body of air which remains over a source region a sufficient length of time assumes certain definite properties in the vertical, particularly with respect to temperature and moisture. Once these properties have been attained, equilibrium with respect to the source region is reached, and any further stagnation or movement of the body of air over the source region will not appreciably affect the balanced distribution.* It is clear, however, that any movement of the air mass away from its source region will result in a modification of its properties. For example, if a body of air from northern Canada moves southeastward into the United States, there is bound to be some warming and moistening. Such modifications tend to destroy the homogeneity originally established at the source region. It is obvious that the modification will take place essentially within the lowest layers of the air mass, the upper layers being modified only gradually by means of the indirect processes of mixing with the modified low layers and by radiation chiefly from the surface of the new region over which the air mass is traveling. The theory of air masses as entities is based upon the fact that the variations of any property in the horizontal in an air mass are small compared with the rapid change of properties observed at the boundary between two air masses which come from different source regions. This boundary zone of rapid transition is a *front.*

From the definition of an air mass it is clear that in order to identify sections of a current as belonging to one and the same air mass we must not only know the properties at the source region and the modifications introduced, but we must also deal with observations which are most representative of these properties. The most representative observations are those made by means of upper air soundings; at the surface the most representative observations are those made at elevated and exposed stations. Examples of non-representative observations are those of valley stations, or those greatly affected by the proximity of a lake or perhaps a mountain barrier. In the latter case there might be an appreciable foehn effect.

As an air mass progresses numerous changes take place, brought about by radiation, mixing (turbulent exchange), adiabatic expansion or compression, and condensation or evaporation. Some meteorological elements will remain more constant than others as the air mass moves from point to

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*The processes by which air masses reach homogeneous equilibrium in their source regions are not yet well understood; further discussion of this appears in the Appendix by Prof. Willett on the air mass properties; a recent study by Wexler (*Mo. Wea. Rev.*, April, 1936) discusses the origin of Polar continental air, and J.-J. Jang has studied the formation of tropical marine air in the trade winds (see Bibliog.).—*Ed.*

point. Furthermore, quantities indirectly obtained by calculation from the observations will vary in constancy. The relative degree of constancy of a meteorological quantity within a moving air mass is defined as its *conservatism.*

We are now in a position to test the meteorological elements and the indirectly calculated quantities with respect to their degree of constancy (conservatism) as the air mass moves.

A. Temperature.

The temperature of any given particle of air within a moving air mass is influenced by the following factors:

1. Conduction and mixing.
2. Condensation and evaporation.
3. Expansion and compression (adiabatic changes).
4. Insolation and radiation.

Temperature, particularly at the surface, is so much changed by these factors that it cannot be regarded as a very representative element by which to identify an air mass after it has moved away from the source region.

The effect of condensation may be eliminated by the use of a quantity called *equivalent temperature.** This is defined as the temperature a particle of air would have if it were made to rise adiabatically to the top of the atmosphere in such a manner that all the heat of condensation of the water vapor were added to the air and the sample of dry air were then brought back to its original pressure. Numerically this is not much different from the temperature the mass of air would have if all its moisture were made to condense and the heat given off by condensation were added to the remaining dry air. Any change in the moisture content of the air mass by condensation will not affect the equivalent temperature of a particle, since the quantity of moisture subtracted by condensation involves a certain loss or gain of heat which is implied in the definition of equivalent temperature.

Evaporation does not greatly change the equivalent temperature; but this is not an important limitation for our practical purpose*.

Changes in the temperature of a particle by means of expansion or compression (adiabatic changes) may be eliminated by the use of the *potential temperature.* Potential temperature is that temperature a parcel of air would have if it were brought adiabatically to a pressure of 1000 mb. If a dry particle were vertically displaced it would warm or cool at the dry adiabatic rate, and therefore its actual temperature would differ from its potential temperature by about 1 C deg. per 100 meters from the level where the pressure is 1000 mb. This holds only in the event that the air particle remains unsaturated, for as soon as it becomes saturated, latent heat of condensation is realized and the particle no longer follows the dry adiabat, but the saturated adiabat. This was pointed out in the first article of this series. During ascent, therefore, the potential temperature of the saturated particle will increase by virtue of the latent heat of condensation.

The most conservative thermal quantity is the *equivalent-potential temperature.* This is the temperature the chosen air particle would have if it were brought adiabatically to the top of the atmosphere so that along its route all the moisture were condensed (and precipitated), the latent heat of condensation being given to the air, and then the remaining dry

*Footnote on equivalent-potential temperature on next page also applies to the equivalent temperature.—*R. G. S.*

sample of air compressed to a pressure of 1000 mb. The equivalent-potential temperature may also be defined as the potential temperature of the equivalent temperature; that is, it can be determined by finding the equivalent temperature, then reducing this adiabatically to a pressure of 1000 mb. The equivalent-potential temperature combines the processes involved in the definition of the potential and the equivalent temperature; hence it is independent of any effects due to expansion or compression as well as condensation*. If we deal with the equivalent-potential temperature of a particle then the only processes which change its value are (1) conduction and mixing, (2) evaporation,* and (4) insolation and radiation. (1) and (4) obviously have their maximum effect in the surface layers, and are probably much less important at higher levels. Therefore the thermal properties of an air mass are far more conservative at upper levels than in the low layers. Consequently it is best to use upper air data rather than surface observations as criteria for the determination and identification of air masses.

B. Lapse Rate.

The lapse rate associated with an air mass is frequently a good index for identification purposes. As in the case of thermal quantities, the lapse rate is much more conservative at upper levels. In the surface layers the lapse rate will be found to vary appreciably from day to day, and from nighttime to daytime. These effects are mainly the result of radiation and turbulence. For example, in the early morning hours there is apt to be a ground inversion, while during the afternoon an adiabatic lapse rate may extend up to about 800 meters. At higher levels surface effects are comparatively small, but there are times when the lapse rate aloft changes because of extensive rising or sinking movements. In spite of variations in the lapse rate in the surface layers, the lapse rate is often indicative of the trajectory of the air current, for when moving over a cold surface the low layers of the current tend to become more stable, whereas when constantly moving over a warmer surface the lapse rate becomes s t e e p e r. Characteristic types of clouds are commonly associated with certain lapse rates.

C. Humidity

The use of aerological material is becoming increasingly important for the investigation of synoptic meteorological problems, especially when air mass and frontal methods are applied. The correct identification of individual air masses is greatly facilitated by the utilization of certain quanti-

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*The equivalent-potential temperature defined by Rossby is not perfectly conservative for an evaporating process, such as when falling precipitation is evaporated into dry air or a fog is dissolved, though the effect does not change the equiv.-pot. temp. greatly. However, as Bleeker has pointed out (*Q. Jn. Roy. Met. Soc.*, Oct. 1939), misleading conclusions as to air-mass identity and movements can be made from the e.-p. temp. if evaporation takes place. The thermodynamic properties which are conservative for evaporation are not conservative for adiabatic changes and condensation; there is, in fact, no element which is conservative for all conceivable changes of the air. In Germany Rossby's equivalent potential temperature is also known as the "pseudo-potentielle Temperatur" (Stüve). Another quantity differing from Rossby's but sometimes referred to as the "equivalent potential temperature" (Normand, 1921) and "equi-potentielle Temperatur" (Robitzsch, 1928), is conservative for evaporation but not for adiabatic processes and condensation; it is defined as the temperature that would result by condensing out all the water vapor *but at a constant pressure of* 1000 *mb.* This of course does not give the same result as when an air mass is raised *adiabatically* to lower temperatures (*and lower pressures*), as in Rossby's definition. Strictly speaking the Normand-Robitzsch quantity is an *isobaric equivalent-potential temperature,* whereas the Stüve-Rossby quantity is a *dry-potential adiabatic equivalent temperature,* and the corresponding equivalent temperatures should be distinguished likewise. Prof. Petterssen in his new book has introduced the terms *pseudo-equivalent,* and *pseudo-wet-bulb potential temp.*, etc., for the adiabatic quantities. Such terminology is too cumbersome in practice but it is well to understand the differences, as the literature and practice are very loose and misleading about these concepts.—*R. G. S.*

ties which remain about constant as the air mass moves from point to point. Chief among such conservative quantities is the specific humidity—a term which is seldom defined in the elementary American textbooks on meteorology, and which, prior to the adoption of air-mass analysis, was almost completely ignored by synoptic meteorologists. In consideration of this it will be of interest to review, qualitatively and in brief fashion, the definitions of hygrometric terms and their relative degree of constancy when an unsaturated air particle containing some water vapor is subjected to vertical displacement.

The vertical displacement of an air particle in the atmosphere brings about changes in the pressure, since at any point the pressure within and surrounding the particle must be equal. If it rises work must be done in expanding against the decreasing pressure; if it descends work is done upon it. This work is realized as a change of temperature of the vertically moving particle. If no heat is added to or substracted from the moving element, in other words if the particle remans thermally insulated, the process is termed adiabatic. The rate of temperature change with altitude of an unsaturated air particle due to these adiabatic changes is very nearly constant (about 1° C. per 100 m.), although there are small variations in this constant due to the water vapor content and the temperature of the surrounding air in which the particle is moving. The first of these corrections, that for the presence of water vapor, becomes manifest when one considers the difference in the specific heats of dry air and of water vapor. The correction, however, is very small, since the percentage of water vapor within a unit mass of air rarely exceeds a few percent of the total. The correction for the temperature of the surrounding air is also relatively small. Accordingly the dry air is the governing factor in the adiabatic rate of cooling of unsaturated air. Practically this amounts to saying that the volume occupied by the vapor and its temperature at any stage are essentially the same as that of the dry air mass with which the vapor is associated.

This facilitates a discussion of the variations of hygrometric quantities in the case of ascending and descending motion.

1. *Vapor pressure* (e). This term represents the partial pressure exerted by the water vapor molecules in the atmosphere. As an unsaturated particle of air is brought upward it expands. This expansion is effected by a change in pressure and a change in temperature which it has been shown is due to the pressure change. The adiabatic temperature change acts volumetrically in the opposite sense to the pressure change, since decreasing temperature causes a contraction of volume. The net effect of these two opposing factors is an increase of volume, so that the original amount of water vapor present within the sample of air must fill a larger space. Hence the vapor pressure (e) decreases with adiabatic expansion.

2. *Relative humidity* (f). This term is defined as the ratio of the actual vapor pressure (e) and the maximum vapor pressure (e_m) possible at the same temperature. It should be noted that this maximum vapor pressure, e_m, is a function of temperature only and has nothing to do with atmospheric pressure. An unsaturated air particle rising rapidly through the atmosphere must cool nearly adiabatically and therefore the maximum possible water vapor pressure, e_m,

must become smaller and smaller. Since the vapor pressure within the air particle (e) is also decreasing, the variation in the relative humidity (e/e_m) really depends upon the rate of change of both these quantities. It so happens that the rate of decrease in the denominator, that is, the falling off of e_m due to adiabatic cooling, is much greater than the rate of decrease in the vapor pressure, the numerator. Thus the percentual value of the fraction, defined as the relative humidity, must increase as the particle rises.†

3. *Absolute humidity.* This quantity is defined as the mass of water vapor present in a given volume, or in other words the density of the water vapor. Again subjecting the air particle to adiabatic lifting it is obvious that the volume of air is increasing as the particle rises. Yet the number of water vapor molecules within the sample of air remains constant, so that the density of the vapor, or the absolute humidity, must decrease with adiabatic expansion.

4. *Specific humidity* (q). This quantity is defined as the mass of water vapor present in a unit mass of air. The unit mass of air is considered as being made up of the usual gas mixture plus the water vapor. The amount of water vapor, however, is negligibly small when compared with the mass of dry air with which it is associated. Hence the mixing ratio, defined as the mass of water vapor in a unit mass of *dry* air, and convenient for certain theoretical calculations is very nearly numerically equivalent to the specific humidity. Indeed errors in hygrometry in aerological soundings are generally much greater than the difference between the two quantities.

As the unsaturated air particle ascends adiabatically, both the mass of water vapor and the total mass of air remain constant. Hence the specific humidity, which depends on these two masses, also remains unchanged. Thus the air particle has the same specific humidity in spite of its rise.

The significance of this constancy becomes apparent when one considers the active overrunning of warm air over a cold wedge (a warm front), forced vertical displacement by an advancing wedge of cold air, or the sinking of air layers within a cold air mass or in a stagnating anticyclone. In all these cases vertical movements bring about changes in vapor pressure, relative humidity, and absolute humidity of the vertically moving particle. The specific humidity, however, does not change, providing no moisture is added to or subtracted from the air through precipitation, evaporation, or turbulent exchange.

At the surface of the earth the moisture content depends upon the synoptic air mass present and the modification which it has undergone during its history. Since the modification is generally greatest in the lowest layers of the atmosphere it follows that the meteorological elements within these layers are less conservative than at higher levels. Nevertheless, the proximity of our American air-mass source regions, and the marked difference in our air-mass properties* bring about considerable contrast in the elements even within the surface layers. The specific hu-

†*Note:* In U. S. the dew-point temperature is reported in the airways hourly observations, and also in the six-hourly reports from first-order stations, for the humidity element; it is not strictly conservative when changes or differences of pressure are involved, but is useful for local comparisons and for fog forecasting.—*R. G. S.*

*For a thorough treatment of these properties see "American Air Mass Properties" by H. C. Willett, Papers in Physical Oceanography and Meteorology, Mass. Inst. of Tech. and Woods Hole Oceanographic Institution. Vol. 2, No. 2, 1933 now out of print but largely reprinted in back of this booklet. Articles on air mass properties in other countries are listed in the BIBLIOGRAPHY.

midity at the surface offers a worthwhile aid to the identification of individual air masses as well as in the upper air. In winter the specific humidity at stations on either side of a well marked front may differ by ten grams per kilogram of air. Although it is true that with large specific humidity contrasts at a frontal zone one generally finds good temperature contrasts, there are many cases in which the temperature immediately after a front passage changes but slightly with the wind shift, yet the specific humidity changes considerably. This is particularly true in summer in connection with air masses of continental characteristics which displace maritime air masses. As an example of this type of front a case may be cited in which a Tropical maritime air mass over New England was displaced by an older transitional air mass originally of Polar Pacific origin. At Boston the specific humidity within the tropical air mass averaged about fifteen grams per kilogram. With a shift of wind from south-southwest to southwest the specific humidity fell fairly rapidly to ten grams per kilogram, even though the temperature rose a few degrees Fahrenheit. It cannot be doubted that the addition of the specific humidity to the data on our daily weather maps would facilitate the analysis and might be the deciding factor in the correct placement of a front when other indications are not pronounced.

The numerical value of the specific humidity is readily calculated by use of the approximate formula

$$q = 622\, e/p$$

where q represents the specific humidity expressed in grams per kilogram; e, the existent vapor pressure (obtained through the relative humidity and temperature observations); and p, the total atmospheric pressure. The units for e and p may be chosen arbitrarily (providing both are expressed in the same unit) since they constitute a ratio.

D. Condensation Forms.

The type of cloud formation is largely a result of the vertical distribution of temperature (the lapse rate) and moisture. Since both these quantities are fairly conservative, it follows that certain condensation forms are more or less characteristic of each type of air mass. Care must be exercised in differentiating between clouds formed within an air mass and those formed by the interaction at the front between two different air masses. In addition, local condensation forms, such as ground fogs, must be eliminated.

E. Visibility.

The visibility in the lower layers of the atmosphere is generally an indication of the lapse rate therein. If the lapse rate is stable, then smoke and dust tend to remain close to the surface; if the lapse rate is steep then vertical motion is easily possible and the foreign matter diluted by being mixed throughout a layer of considerable thickness, thereby increasing the visibility in the surface layers. In cold masses which are moving over a much warmer surface a steep lapse rate is soon established and visibilities become good. On the other hand, when a warm current moves over a much colder surface the marked stability keeps the foreign matter concentrated in the lower layers making the visibility poor. As an index of the air mass present, visibility must be used with considerable care, since there are numerous factors other than turbulence affecting visibility.

F. Wind Direction and Velocity.

Wind direction and velocity are in themselves not very conservative elements. Polar air masses are frequently observed with winds having a southerly component, and tropical air masses not infrequently have sections in which the wind may blow from the northwest. This is true particularly in upper levels. In the placement of fronts, however, winds are very important. Some of the fundamental concepts of this phase of the problem will be taken up in a later article in this series.

Of the six elements named above, the thermal and hygrometric quantities are the best indices to be used in following and identifying air masses. Of these, the equivalent-potential temperature is the most conservative, combining the conservative qualities of both the potential temperature and the specific humidity. The above discussion, necessarily brief and sketchy, will enable us to take up the Rossby-diagram in the next article of this series.

III. THE ROSSBY DIAGRAM—PLOTTING ROUTINE

It is of primary importance in synoptic meteorology that air masses be followed as they move from area to area over the earth's surface; the weather at any given locality is, of course, largely dependent upon the type of air mass present and the modification which the body of air has undergone during its history. In the preceding article it was pointed out that the use of representative observations is necessary for the identification of air masses from different source regions. The upper layers of the atmosphere being comparatively free from surface effects, it is best to use data from upper air soundings. It was also shown that the most conservative quantities that can be used for purposes of identification are not the ones directly measurable—temperature and relative humidity—but those indirectly obtained —potential temperature and specific humidity. These two quantities, as will be seen later (Article X), are also used in isentropic analysis.

Realizing the importance of a quantitative method for identifying air masses, Professor C.-G. Rossby, of the Massachusetts Institute of Technology, developed the diagram which bears his name. As might be supposed, the diagram makes use of the most conservative quantities—potential temperature and specific humidity. Potential temperature is the ordinate of the diagram, specific humidity the abscissa. Since the equivalent-potential temperature is a function of potential temperature and specific humidity, another set of lines representing constant equivalent-potential temperature may be constructed on this diagram. The significance of these lines as well as the interpretation of various curves on the Rossby diagram will be discussed later. At present we shall concern ourselves with the mechanical procedure of constructing, on these diagrams, curves representing aerological soundings.

In the first article of this series it was pointed out that, in an adiabatic process, the relation between temperature and pressure is given by the formula:

$$\frac{T_1}{T_2} = \left(\frac{p_1}{p_2} \right)^{0.288}$$

where T_1 is the temperature at the pressure p_1, and T_2 is the temperature the particle assumes at the pressure p_2. If we now specify that the particle, originally at temperature T_1 and pressure p_1 be compressed to 1000 mb pressure we have

$$\frac{T_1}{T_2} = \left(\frac{p_1}{1000} \right)^{0.288}$$

But if the particle is reduced dry

adiabatically to 1000 mb, the temperature, T_2, which it assumes at this pressure is by definition the potential temperature, which we shall call θ. Thus

$$\frac{T_1}{\theta} = \left(\frac{p_1}{1000}\right)^{0.288}$$

$$\text{or } \theta = T_1 \left(\frac{1000}{p_1}\right)^{0.288}$$

From this formula the potential temperature is readily computed. Tables giving the factor $\left(\frac{1000}{p_1}\right)^{0.288}$ may be constructed to facilitate this computation. In practice, however, it is generally more convenient to obtain the potential temperature by means of the adiabatic chart. It is assumed that the reader has access to such a chart.* For convenience of reference, fig. 2 will serve to show the essential features of this diagram. The ordinate consists of the pressure, which, in the latest type of adiabatic chart, is plotted to the power 0.288 (i.e., $p^{0.288}$ is the ordinate). The reason for using $p^{0.288}$ is that when plotted in this fashion against a linear scale of temperature, the lines representing dry adiabats become slanting straight lines (though they converge upward slightly). This is seen from Poisson's equation. In the older Stüve adiabatic chart the pressure is plotted on a logarithmic scale, so that there is a slight curvature to the dry adiabats.*

A little manipulation of Poisson's equation explains the other features of

*Various older forms, such as the Hertz, Neuhoff, and early Stüve diagrams, are still found in many physics and meteorology texts. The more convenient form, Stüve's newer psuedo-adiabatic chart, is now used by most meteorologists and weather services.—*Ed.*

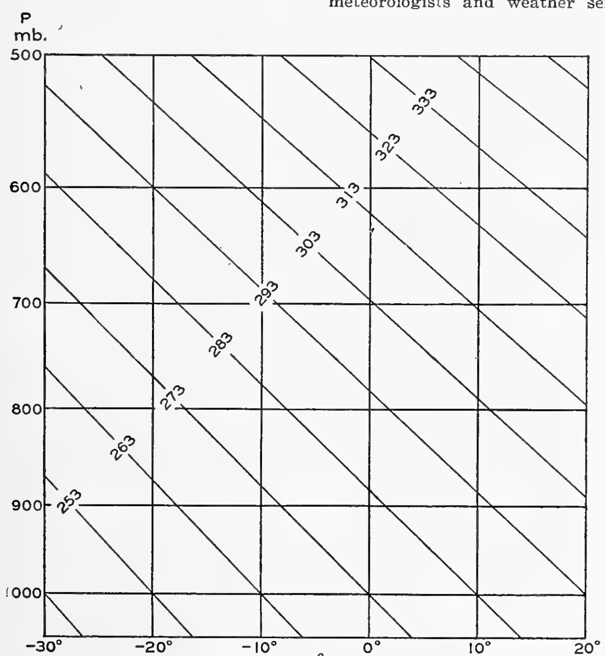

FIG. 2. ADIABATIC CHART (Stüve's $T, p^{0.238}$)

the chart. One can derive that $p_1^{0.288} = T_1 \ (1000)^{0.288}/\theta$ and thus $p_1^{0.288}$ varies only as T_1; θ and $(1000)^{0.288}$ being constants for any individual θ line starting at 1000 mb. Thus T is linear on the abscissae scale, and $p^{0.288}$ is linear on the ordinate scale. Note also that on any given T line the vertical intervals between unit θ values diminish upward, since θ varies as $T/P_1^{0.288}$. The convergence upward of the dry adiabats follows from the relation that $p_1^{0.288}$ varies as T_1/θ, so that as θ increases the slope T_1/θ decreases along the 1000-mb line towards lower values of θ.

[Wet adiabats (= equivalent-potential or wet-bulb temperature isotherms) are often added to the adiabatic chart making it a *pseudo-adiabatic diagram* on which the behavior of saturated particles can also be studied; but this is of no concern to the present discussion. Saturation specific humidity lines may also be added on the $p^{0.288}$ type of chart. These have the advantage of being straight lines but are curved on the log p diagram.]

If the pressure and temperature of any particle of air be known, then one may find its potential temperature with the aid of the adiabatic chart by moving the original point parallel to the dry adiabatic lines until the 1000 mb line is intersected. The temperature at this intersection is the potential temperature. An easier way is to label the slanting lines (the dry adiabats) with the particular potential temperature which remains constant along them. Thus any point on the adiabatic chart which is determined by a pair of values of p and T has one potential temperature, which may be ascertained from the dry adiabats. Potential temperature is practically always expressed in degrees Absolute.

With the help of the adiabatic chart and definitions which have been given, the student should be able to deduce the following important generalizations:

1. *The potential temperature along a dry adiabatic line is constant.*

2. *A decrease in potential temperature with elevation corresponds to a superadiabatic lapse rate.*

3. *The greater the rate of increase in the potential temperature with elevation, the greater the stability.*

It is essential that the reader be thoroughly familiar with the concept of potential temperature in order to appreciate the Rossby diagram.

It should be mentioned here that, strictly speaking, the potential temperature as here defined is not the quantity used as the ordinate in the Rossby diagram, but a very nearly numerically-equivalent quantity known as the partial potential temperature with respect to dry air. The total pressure exerted by air is made up of the pressure of the dry air (i.e., all the gases with the exception of water vapor) plus the pressure exerted by the water vapor. The partial potential temperature with respect to dry air is then defined as the temperature the particle would have if it were reduced adiabatically from the pressure exerted solely by the *dry* air to a pressure of 1000 mb. The algebraic representation of this quantity will serve to clarify the definition. If θ_d represents the partial potential temperature with respect to dry air, p the total pressure of the air, T the temperature of the particle, and e the partial pressure exerted by the water vapor, we have:

$$\theta_d = T \left(\frac{1000}{p-e} \right)^{0.288}$$

This is, of course, similar in form to the formula for the potential temperature, $(p-e)$ replacing the p in the expression for potential temperature. But e is generally very small compared with p, and for this reason θ_d and θ do not differ appreciably. It is doubtful if, in practical meteorological work, it is worth while to obtain θ_d rather than θ. However, if numerical accuracy is desired, the partial potential temperature may be obtained by entering as the ordinate of the point on the adiabatic chart, not the actual pressure, but the total pressure minus the pressure due to the vapor. This graphical process is facilitated by locating the actual point of pressure on the adiabatic chart, then displacing this point vertically upward by the number of millibars of pressure exerted by the vapor.

On one side of the adiabatic chart is generally found a set of curves by means of which the specific humidity may be determined. A more convenient method of computing this quantity will be given below. The reader, however, will find it helpful to discover how to use the specific humidity curves on the adiabatic chart.

It has been pointed out that the specific humidity may be computed from the formula:

$$q = 622 \frac{e}{p}$$

where q is the specific humidity expressed in grams per kilogram of air; e, vapor pressure, and p, the pressure of the air. The units for e and p may be chosen arbitrarily since they constitute a ratio. They are generally expressed in millibars, however, since upper air observations are transmitted in these units. The value of e, the vapor pressure, is readily obtained by multiplying the relative humidity by the saturation vapor pressure at the temperature of the particle. The saturation vapor pressure for different temperatures may be found in most texts on meteorology or physics. Thus if one wishes to find the vapor pressure at the temperature of 10°C and relative humidity of 50%, he finds in the saturation vapor pressure tables the value of 12.28 mb corresponding to 10°C. Since the air is only 50% saturated, the vapor pressure is one-half of 12.28 mb or 6.14 mb. Note that vapor pressure is entirely independent of pressure, being a function of temperature and relative humidity alone. After e has been determined, the specific humidity, q, is calculated by means of the above formula. A slide rule is most convenient for this computation.

It should be also noted that, strictly speaking, the specific humidity is not used as the abscissa of the Rossby diagram, but rather a very nearly equivalent quantity called the mixing ratio. This term is defined as the mass of water vapor per unit mass of perfectly dry (absence of water vapor) air. In algebraic form:

$$\text{mixing ratio } (w) = 622 \frac{e}{p-e}$$

From the above formula it is clear that since e is very small compared with p there will be no appreciable error introduced by the use of q instead of w. In fact, with the present inaccurate method of measuring the relative humidity in the upper atmosphere with the hair hygrograph, it is ridiculous to try to obtain such an accuracy as the difference between the formulae for q and w indicate. The use of the partial potential temperature and mixing ratio rather than potential temperature and specific humidity as coördinates of the Rossby diagram, was made necessary by the construction of the lines of

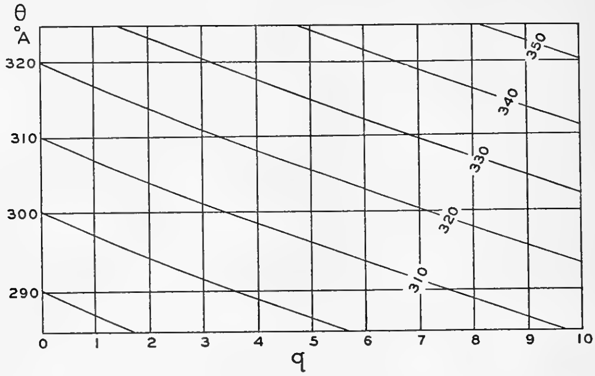

FIG. 3. ROSSBY DIAGRAM IN OUTLINE

equivalent-potential temperature. In practice it is most convenient to use the potential temperature and the specific humidity; hereafter when the coördinates of the Rossby diagram are mentioned, they will be referred to simply as potential temperature and specific humidity, or in symbols as θ and q, with the understanding that the reader is aware of the small differences between them and the partial potential temperature and mixing ratio.

The abscissa of the Rossby diagram, the specific humidity, is a linear scale (see fig. 3); the ordinate, potential temperature, is logarithmic. This construction facilitated the computation of the lines of constant equivalent-potential temperature in the original drawing. The position of any point on the diagram is thus obtained by its values of q and θ. Thus if one wishes to plot an airplane sounding on the Rossby diagram there is a routine involved in computing q and θ. The reports of soundings transmitted over the teletype system or by short-wave radio, contain the altitude in meters (z), the pressure in mb (p), the temperature in Centigrade (T), and the relative humidity in percent (f). The following working table is suggested:

Station and date of sounding

z	p	T	f	e	q	θ	Remarks

After q and θ are obtained the diagram is easily plotted.

Abbreviations for the names of the stations and the dates of the soundings may be attached to the ends of the curves, and several curves may be drawn on one diagram, providing they do not interfere too much with one another. Dotted lines, broken lines, and the like help to simplify the differentiation among curves.

The lines of constant equivalent-potential temperature, which are determined by means of the θ and q scales, are those curves sloping downward from left to right.

The interpretation of the various types of curves that appear from plotting soundings on the Rossby diagram will be treated in the next article in this series.

Too much emphasis cannot be placed on the importance of the practical experience with the methods discussed in these articles. It is mainly by experience that the student learns to understand clearly and really make use of them.

USE OF THE ROSSBY-DIAGRAM AS A NOMOGRAM FOR FINDING θ AND q

Professor Rossby has also suggested another method of obtaining the potential temperature and moisture content directly from the equivalent-potential temperature diagram which is presented at the rear of Rossby's original paper.[2] This particular diagram differs from that described above (see fig. 3) in that it contains, in addition to the θ, w, and θ_E (equivalent-potential temperature) lines, the isobars and isotherms for the condensation level. With this particular diagram (and with one operation in multiplication, which also may be done graphically) θ and w may be readily evaluated from the values of p, T, and f obtained in an airplane sounding. Dr. Byers' description of this method appears below. It is undoubtedly a time saving process, although one may find some difficulty in working with a chart which has five sets of intersecting lines. Anyone inexperienced with the Rossby diagram will find it better to practice first the basic methods presented above until he has a good understanding of the fundamental quantities and operations.—*J. N.*

[The Rossby diagram or equivalent-potential temperature chart is available in a form which gives the pressures and temperatures at the condensation point of the air after an adiabatic expansion from the observed conditions of potential temperature and moisture content. A determination in a reverse manner makes it possible to evaluate the potential temperature and moisture content from the ordinary observed pressure, temperature and relative humidity data (See Plate I, in reference cited in footnote 2, p. 16).

The assumption is made at first that the air has just reached saturation. The point of intersection of the actual temperature line (under the assumption this is also the temperature of the condensation point) with the line for the given pressure coincides with the point of intersection of the lines of actual potential temperature and *saturation* moisture content. Multiplying the latter by the relative humidity gives the existing moisture content. Since the potential temperature in an adiabatic expansion is constant until saturation is reached, no correction is needed for this quantity.

The graphical method suggested by Prof. Rossby is as follows: With the observed temperature expressed in absolute degrees, enter the diagram along the corresponding red temperature line until the observed pressure line is intersected. Read the potential temperature at the point of intersection from the black lines which run horizontally across the diagram and the saturation moisture content from the black vertical lines. Multiply the latter value by the relative humidity. For this operation, a graph can be made consisting of saturation moisture content and actual moisture content as coördinates, with relative humidity lines crossing diagonally.

Example: Pressure 860 mb. Temp. 3°C., or 276°A. Rel. Hum. 80%.

Follow red line of 276° to 860 mb, where potential temperature of 288.1 and moisture content at saturation of 5.5 g/kg are indicated. Taking 80% of this latter value, we find the actual moisture content, which is 4.4 g/kg. —*H. R. Byers*].

Millar of the Canadian Meteorological Service has designed another

[2]C.-G. Rossby, Thermodynamics applied to air mass analysis, Mass. Inst. of Tech., Meteorological Papers, vol. 1, no. 3. 41 pp., 1932.

nomogram for q and θ, which is shown in Fig. 4.—*R. G. S.*

Note.—Beginning in the summer of 1938 the U.S. Weather Bureau ceased regularly transmitting over its telegraph and teletype circuits the potential temperature (θ) for each significant level of the daily aerological soundings. It became necessary therefore to compute all the derived values needed for plotting various energy and thermodynamic diagrams or θ_E lapse-rates, etc. However, the Bureau at the same time published for its offices a new type of adiabatic chart base (*"Upper Air Map C"*) which contains, besides the fundamental grid of ordinary temperature *vs* pressure, only the slanting curves of the wet (saturation) adiabats; but on the margins are numbered marks to indicate where the lines of potential temperature (straight), of saturation specific humidity (also straight) and of kilometers of elevation above the sea, would intersect. By using a ruler, or better still, a celluloid or glass transparent plate with sets of lines for potential temperature and saturation specific-humidity drawn to scale on it, one can quickly evaluate these figures from the base chart once the sounding is plotted thereon in the ordinary terms of T, p, and q. Thus the new chart serves as a convenient nomogram. It supplants the temperatures *vs* height aerological plotting chart which has heretofore been used for years in the Weather Bureau forecast offices and elsewhere. Owing to the small size of this chart it is not suitable for accurate work and will be supplanted by a new and larger design. The large standard Stüve adiabatic chart has long been available but used chiefly at aerological stations; it is too large for ready comparison of many soundings at once, as the forecaster must do in some countries.—*Editor.*

Modified Rossby-Diagrams

A number of variations have been proposed and some are in use. They have some advantages in convenience but little if any fundamental ones. Any diagram with two conservative elements among its coordinates would essentially still be a Rossby diagram. For θ_D Brunt (Phys. and Dyn. Met., 2nd ed., p. 92) has suggested substituting the (saturated) potential wet-bulb temperature and Hewson has analyzed situations in that manner (see Bibliography). However, Bleeker points out that this quantity is not conservative if any evaporation (of falling rain, or by subsidence) takes place, when it may lead to erroneous conclusions as to movement and identity of air masses. Rossby's equivalent-potential temperature is somewhat less conservative for an evaporation process than the adiabatic wet-bulb potential temperature. However, there is no single index which is conservative for both adiabatic processes and evaporation, and since evaporation does not often seriously invalidate the usual interpretation of the Rossby diagram the adiabatic wet-bulb and equivalent-potential temperatures are both equally the best conservative elements available.

Clark of California Institute of Technology revised the Rossby-diagram by arranging the θ_E isotherms vertically and θ_D lines horizontally and plotting curved lines of constant mixing ratio and isobars of the condensation level upon the chart. The characteristic curves will vary in slope through wider angles than on the original diagram, so that different stability conditions are more easily recognized at a glance (described in Taylor's Aeronautical Meteorology, p. 60.).

Arakawa has published in Japan a new form of the Rossby-diagram having for ordinates the potential temperature on a linear scale instead of partial potential temperature on a logarithmic scale, and having equivalent-potential temperature lines which are calculated on the basis of convection in saturated air with the heat of fusion taken into account; also the diagram is extended to over 20 g/kg mixing ratio so that it can be used for extremely humid tropical conditions. The advantage of these coordinates is that the effect of mixing of two air masses can also be determined, since the potential temperature of a mixture is equal to the mean of the potential temperatures of the original air masses (*Bull. Amer. Met. Soc.*, March 1940, p. 111).

Finally, the Refsdal "Aerogram" should be mentioned here since it combines the properties of the adiabatic chart and the Rossby diagram and the tephigram in one, though it is rather too complicated for the elementary student (see:—*Geofysiske Publik.*, Vol. XI, No. 13; *Meteorol. Zeit.*, Jan. 1935, p. 1; *Bull. Amer. Met. Soc.*, Jan. 1940, p. 1).

(See also "A Note on Estimating Conditional and Convective Instability from the Wet-bulb Curve", at the end of Article VIII in this booklet.)

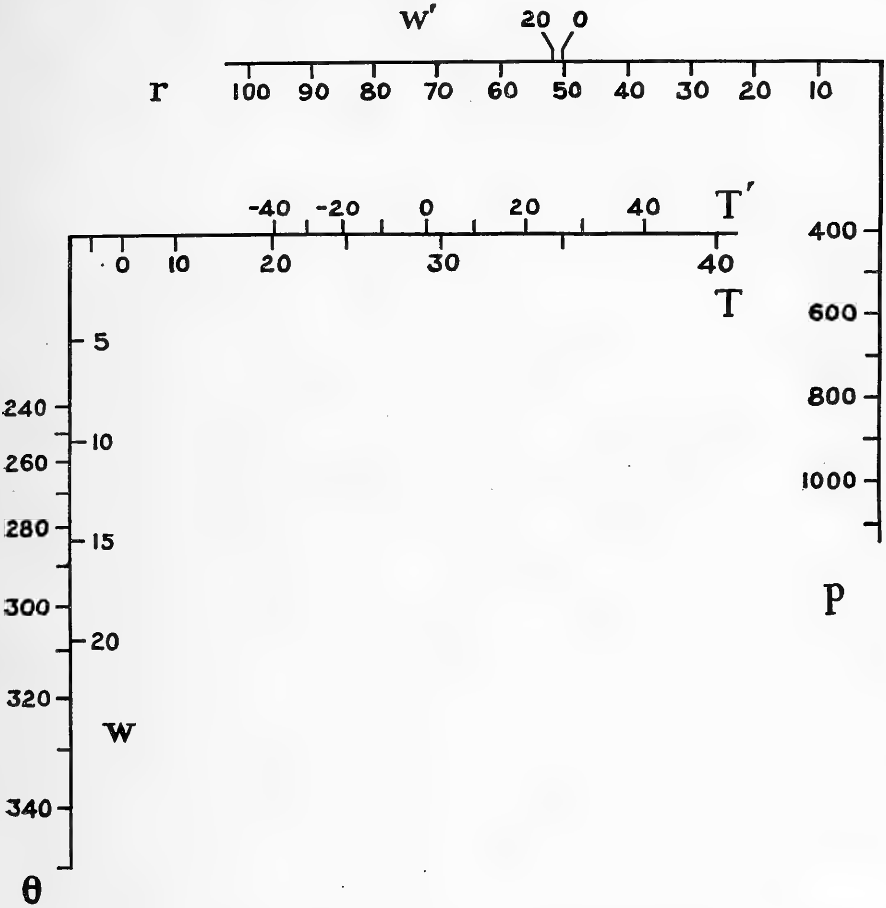

FIG. 4. THE MILLAR NOMOGRAM FOR RAPID EVALUATION OF w AND θ, SHOWING MAIN DIVISIONS ONLY.

DIRECTIONS: To obtain w, lay a ruler across the p and r scales, setting the edge on the pressure p, and the percent relative humidity r. Then, holding the ruler rigid, hold a 90° set square against the ruler, and slide the square along until the edge is set at the temperature on the T scale. The corresponding value of w will then be read against the square on the w scale. To obtain θ, proceed in the same way, but setting on the p, w', and T' scales. For w', use the value of w previously obtained. (For further details see *Bull. Amer. Met. Soc.*, Oct. 1935, p. 229; or June-July, 1936, p. 18).

IV. THE ROSSBY DIAGRAM—INTERPRETATION

In the preceding article the mechanics of plotting the Rossby diagram were discussed, and it may be said here that the curve joining the individual points is called the characteristic curve for the air column. The present article deals with the significance of various curves on the diagram and the changes in appearance of these curves as the more common atmospheric processes, such as vertical displacement, occur. At this point it is only fair to remind the reader that the following discussion is, from the standpoint of a quantitative analysis, rather unsatisfactory in that it is an attempt to describe processes which are better expressed in more exact mathematical terms. Yet it is hoped that the following will serve as an outline for those who have a limited knowledge of mathematics and physics.

In the Rossby diagram the characteristic curve for an unsaturated particle of air, A (see fig. 5) having a given potential temperature (θ) and specific humidity (q) and being displaced vertically, is represented as a point. Owing to the fact that its potential temperature must remain constant, the point A cannot be displaced from the horizontal line representing its potential temperature. Furthermore, the specific humidity remains constant during the adiabatic process with unsaturated air, and thus the point A cannot be displaced from the vertical line representing constant specific humidity. It is evidènt, then, that the characteristic curve of a dry particle of air which is undergoing adiabatic transformation reduces to a point on the diagram. For a given element of air this point will be represented by the same coördinates (θ and q) until the level of condensation has been reached. From this level on it is assumed that the water produced by condensation drops out immediately—in other words, the process is considered pseudoadiabatic (*pseudo*, since there is a small amount of heat removed from the air by the falling water). From the definition of equivalent-potential temperature (θ_E) it is clear that the characteristic curve of the saturated mass of rising

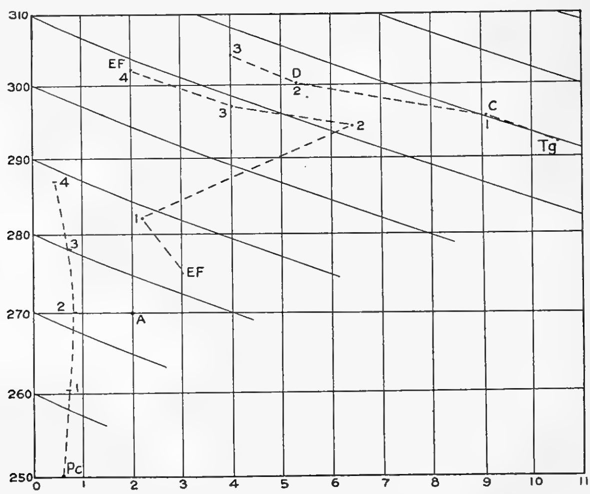

FIG. 5. CHARACTERISTIC CURVES ON THE ROSSBY DIAGRAM.

air is a line of constant equivalent-potential temperature. This amounts to saying that the lines of constant θ_E are also the saturation adiabats. In a rough fashion it is evident that the θ_E lines must slope as they do; the potential temperature increasing in the saturated air particle because of the realized (latent) heat of condensation, and the specific humidity falling because of the condensation and removal of the liquid water.

The fact that an adiabatic process with unsaturated air is represented by a point makes the Rossby diagram particularly adaptable to modern synoptic analysis. A thin layer of air having a uniform distribution of temperature and moisture may be represented on the diagram as a straight line. Let this line be CD in fig. 5. If the entire layer CD be raised or lowered the temperature and relative humidity of each particle of air within the layer will be changed because of the adiabatic expansion or compression. But the potential temperature and specific humidity of each particle of air will remain unchanged, so long as condensation or evaporation do not take place, and for this reason the layer, compressed or expanded, will be represented on the diagram by the same line CD. It is important to note that what has been said refers to one stratum of air, no new strata being introduced or removed during the process. The line element CD may then be considered as the characteristic curve for the given layer. If an entire aërological sounding is plotted on the Rossby diagram, it may be considered as the characteristic curve of the air column through which the sounding has been made.

In the definition of an air mass, horizontal homogeneity was stressed. Soundings, then, made at different places within a source region and remaining within the same air mass should exhibit nearly identical characteristics; the characteristic curves should be similar. Frequently the air masses, even at the source regions and particularly after travelling some distance, are subjected to forces which lead to appreciable vertical displacements. When this occurs unequally in different sections of the air mass the homogeneity with respect to level tends to be destroyed. The surfaces of constant potential temperature and constant specific humidity become curved instead of horizontal, and plots of temperature or moisture against elevation appear markedly dissimilar. It follows that it is difficult to identify and follow air masses by means of these diagrams. The characteristic curves on the Rossby diagram, on the other hand, will have overlapping parts for the same column of air regardless of the extent of the expansion or compression, providing no condensation, evaporation, or introduction of a new air mass has taken place. The reader may refer to papers published by the Meteorology Course of the Massachusetts Institute of Technology and to Harvard Meteorological Studies, No. 2, for examples of this characteristic quality for various air masses on the Rossby diagram.

Fig. 5 illustrates characteristic winter curves for two air masses: one having its source over Northern Canada (Pc—Polar Canadian) the other from the Gulf of Mexico (Tg—Tropical Gulf).* The numbers, beside points on the curve, represent elevations in kilometers above sea level. The Pc curve exhibits the pronounced coldness, dryness, and stab-

*For description of these air masses see the article by Prof. Willett in the back of this booklet.

ility of the Polar Canadian air. The TG curve shows the warm and moist character of the tropical maritime air.

The Rossby diagram is helpful in the treatment of stability. Thermodynamic diagrams, for the most part, deal with the stability or instability of a particle of air with respect to its surroundings. This is of importance in penetrative convection, such as in cumulus cloud and thunderstorm formation. However, in the more important types of convection, that is, in the case when a layer of warm air of large horizontal extent is forced over an underlying cold air wedge or a mountain range, the classical energy diagrams fail to give a measure of the potential energy available in the layer. It is here that the Rossby diagram is invaluable. It is true, however, that in regard to particle stability the classical diagrams are more useful than the Rossby diagram, but even in this type of stability it is not difficult to apply the equivalent-potential temperature diagram of Rossby. This may be done with the aid of the lines of equal potential temperature (the horizontal lines) and the elevations of the points which may be conveniently indicated by figures beside the individual points of the sounding. Thus if there is no increase in potential temperature through a layer, the lapse rate is equal to the dry adiabat, or 1 C deg. per 100 m. If there is an increase in potential temperature of 10 deg. in 1000 m the lapse rate is isothermal. In general, the greater the increase in potential temperature with elevation the greater the stability. If the potential temperature decreases with elevation, the lapse rate is superadiabatic.

The second type of stability, that within a layer which is being displaced vertically, is best treated by the Rossby diagram. From the slope of the characteristic curve relative to the slope of the lines of constant θ_E, one can determine whether the layer in question is convectively unstable (sometimes called potentially unstable). This condition is defined as one in which the equivalent-potential temperature decreases with elevation, and is indicated on the Rossby diagram by a line which possesses a slope between those of the lines of constant θ_E and constant θ. In other words the line representing convective instability is one which lies between the potential and the equivalent-potential isotherms. The importance of this particular distribution of temperature and moisture lies in the fact that lifting of the entire layer brings about a more unstable condition, in fact, if the layer is lifted sufficiently it will eventually become unstable with respect to dry (unsaturated) air. The following illustration will serve to show this increasing instability. Suppose we have the layer CD, which is by definition convectively unstable, having the equivalent-potential temperature 320°A at the base of the layer and 315°A at the top. If the layer is now lifted pseudo-adiabatically to the top of the atmosphere, the base of the layer will have an equivalent-potential temperature greater than that at the top of the layer by 5 C deg., the additional heat being supplied to the bottom of the layer by the extra heat of condensation given to it by the greater amount of water vapor at the base compared with the top of the layer. This decrease of potential temperature, of course, denotes instability with respect to dry air. If the temperatures at the top and base of the layer were taken at successive intervals in the ascent of the stratum the difference in temperature would be seen to become greater and greater. Thus any layer of air which was originally convectively unstable, when subjected to lifting, will acquire a steeper lapse rate, eventually reaching a state of instability with respect to dry air.

Inspection of fig. 5 will reveal that a decrease in θ_E with elevation may be brought about in different ways. For example, it may be due to a rapid drop in temperature with elevation (that is, a small change in potential temperature), a rapid decrease in the moisture content with elevation, or of course a combination of both. Reference to the relative humidity distribution and lapse rate is sufficient to indicate to the synoptic meteorologist whether the convective instability is due to the moisture or to the temperature distribution. In this connection it is well to note that convective instability may or may not carry with it conditional instability. Similarly, conditional instability is not necessarily accompanied by convective instability.

The idea of convective instability may perhaps be made clearer by dealing with a special case in which the base of a stratum is nearly saturated, while the top is very dry. Lifting of this layer will lead to saturation of the base long before the upper layer. Thus the lower part of the layer cools at the rate for saturated air at those temperatures, while the upper part of the layer cools at the adiabatic rate for dry air. It is obvious that the base of the layer is becoming warmer and warmer relative to the top of the layer, thereby increasing the lapse rate. Furthermore, even after the entire layer becomes saturated, the base of the layer is receiving more heat of condensation than the top of the layer.

The mere existence of some layer or layers of convective instability within an atmospheric sounding does not mean that the energy stored therein will be released in the form of convection. The meteorologist must consider whether vertical forces brought about by thermal or mechanical convection will come into play sufficient to lift the layer enough to convert the potential energy of the layer into kinetic energy. In this connection reference must be made to the synoptic chart.

If the equivalent-potential temperature increases with elevation the state is one of stability with respect to dry or saturated air, and no adiabatic process performed upon the layer can render it unstable.

Another important use of the Rossby diagram is in distinguishing between temperature inversions which have developed within the same air mass and those which are the result of a warm current overrunning a cold wedge. In the latter case the warm current generally has, level for level, a much higher moisture content than the cold current. For example, let us suppose that we are dealing with a wedge of cold air which had its source over northern Canada while a current of warm moist air from the Gulf of Mexico is overrunning it. A sounding made through the cold wedge of air into the warm current would then show a rapid increase of moisture and potential temperature. The rapid increase is shown in the Rossby diagram (curve EF, fig. 5) where the wedge-like curve shows a maximum specific humidity aloft. In one and the same air mass the specific humidity generally falls off gradually. This would be expected in view of the fact that all atmospheric water vapor originates at the earth's surface. On the other hand, soundings obtained within one and the same current of air show no pronounced wedge-like curve, but instead a continued decrease in the moisture content.

The method of making use of the Rossby diagram in practical meteorological work is best grasped by studying synoptic discussions in which the diagrams are used, and if possible, making daily use of them in conjunction with the analyzed weather maps.*

Note on present use of the Rossby and other diagrams.—The Rossby diagram is no longer so extensively used in *daily routine* weather analysis in the United States as some years ago. (The use of tephigrams and emagrams likewise seems to have decreased). The reason for this appears to be in part the lack of time available for plotting additional charts, but more importantly the increasing experience of the analysts which permits them to recognize the air masses and stability conditions from an inspection of the surface map and of the aerological soundings plotted on adiabatic charts or against height. Time available to forecasters is so limited that the introduction of additional charts is resisted by them except where some very indispensable advantage can be readily demonstrated. All synoptic meteorologists agree, however, that some such diagram as Rossby's should be a familiar tool and that it is invaluable for teaching principles to students and for research. The diagram still finds much use as a nomogram for finding θ, θ_E, etc., and other thermodynamic computations. For identifying convective instability it is only necessary to have a table of the θ_E lapse rates for the sounding. In the note on wet-bulb temperature appended to Article VIII is indicated how all types of stability and energy distribution can be recognized on the tephigram or emagram alone, thus obviating the necessity of plotting two or three charts. Even the Stüve pseudo-adiabatic chart, which is usually plotted anyway in all weather services, can be adapted in a crude way for all the purposes attributed to other charts and since time is at a premium there is a tendency for American forecasters to depend on it almost exclusively. In a tropical country the tephigram is probably the most convenient all-purpose chart (see Article VIII). The Refsdal "Aerogram" is designed to combine features of all other diagrams and is being used in some services. Its inconveniently small scale and the complications of many overlapping coordinate grid-lines do not commend it for students, nor for speed and simplicity.—*R. G. S.*

V. ELEMENTS OF FRONTAL STRUCTURE—THE WARM FRONT

Anyone who has worked with a close network of surface observations clearly recognizes the existence of zones in which the rate of change of the commonly observed meteorological elements, e.g., temperature, is comparatively large. These zones of rapid transition of the elements are the "fronts" of the Norwegian system of weather analysis. In the general treatment of air masses it was pointed out that large-scale air currents, in spite of a long trajectory, tend to retain their original properties, this being particularly true at upper levels. It follows, then, that as two air masses, one from polar regions and the other from tropical regions, converge, there must be a zone of separation between the two currents. The methods of air mass and frontal analysis are primarily based upon the postulate that the zone of separation between currents of air possessing different properties may be treated as a surface of discontinuity. Atmospheric discontinuities, to be sure, are not absolute mathematically abrupt transitions, but rather are zones in which the change in the elements is far more rapid than within the air masses on either side of the front.

These discontinuities are not purely surface phenomena; they are surfaces which generally slope backward (or forward) over the cold air. If an ideal case is assumed in which a warm current of air flows side by side but in opposite direction to a cold current the following formula

*See synoptic papers by Willett, Emmons, Namias, and Byers, 1932-1936, in Bibliography. —*Ed.*

gives the slope of the surface of separation between the two currents:

$$(1)\quad \tan \beta = \frac{l}{g} \frac{(T_2v_1 - T_1v_2)}{(T_2 - T_1)}$$

where $\tan \beta$ is the slope of the front, l the deflective force due to the earth's rotation, g the acceleration of gravity, T_1 and T_2 the temperatures of the two currents, and v_1 and v_2 the corresponding velocities. From this formula we see that, other factors remaining the same, the slope of this ideal stationary front becomes greater as the difference in temperature between the two air masses (involving $T_2 - T_1$ in the denominator) becomes smaller. Also, an increasing difference in the velocities on either side of the front, other things being the same, requires an increasing slope.

In the development of the above expression, several assumptions are made which, though probably never rigorously attained in Nature, are frequently approximated, e.g., the air flow on both sides of the front is assumed to possess no curvature. Slow moving fronts often approach this linear character.

The conditions necessary for equilibrium of two air currents which are flowing side by side are, then, that the cold air must underlie the warm air in the form of a wedge and must flow, in the northern hemisphere, to the right of an observer looking from the colder into the warmer air mass.

If, in the atmosphere, these surfaces of discontinuity remained stationary under the conditions of equilibrium outlined above, there would be little or no change in the weather, for weather changes are primarily the result of frontal movements and air mass interactions. Complete equilibrium within the earth's atmosphere is never reached. In other words, frontal surfaces are continually undergoing some modification; perhaps they are becoming steeper, perhaps accelerating, or perhaps undergoing various transformations simultaneously. The reasons for these deviations from ideal stationary conditions are beyond the scope of this series of articles. It can be seen, however, that if a wedge of cold air is too steep, that is, if it exceeds the equilibrium value of slope given by (1), it will tend to flatten out, the colder air spreading out underneath the warmer. This leads to vertically upward components in the warm air ahead of the front which, in turn, may lead to adiabatic cooling sufficient to form cloud. Likewise, a slope which is not steep enough will become steeper. In this case a downward component is established within the overlying warm air. Once the stationary equilibrium of a front is disturbed so that, e.g., there is convergence of wind flow, the warm air is forced to ride over the underlying cold wedge. Then most of the upward component is more likely a result of this convergent flow rather than of any change in slope of the front.

Perhaps the most convenient classification of discontinuity surfaces is one based upon the active or passive nature of the vertical component in the warm air above the cold wedge. In the active case the vertical component is a result of processes which are independent of the cold air; in the passive case, the moving cold air brings about forced vertical movements in the overlying warm air.

In this article we shall deal solely with the class of discontinuities in which the warm air possesses an upward component of motion due to convergence into and ascent over an underlying cold wedge of air. These discontinuities are termed warm

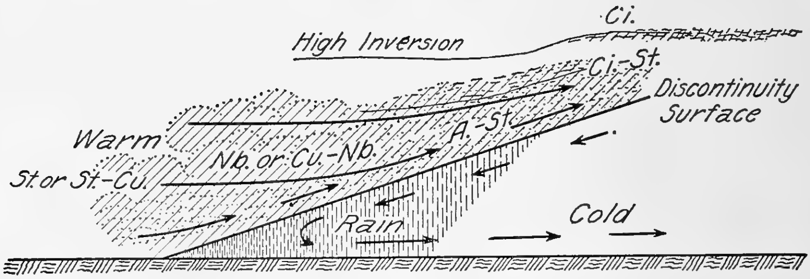

Fig. 6. Idealized Vertical Cross-Section of a Warm Front (After Bjerknes).
(Reproduced from fig. 90, Bull. Nat. Res. Council, No. 79.)

fronts. Cold fronts, aggressive wedges of cold air which force warm air upward, will be taken up in the next article. Discontinuities characterized by a downward component of warm air flow along a surface of cold air are given the name surfaces of subsidence. They also will be treated later.

In the case of a warm front the cold air acts as a barrier to the warm flow, the resulting vertical displacement leading to the formation of clouds and perhaps precipitation. Figure 6 shows the characteristic circulation, cloud forms, and precipitation area generally associated with a well-defined warm front. In the eastern part of the United States the over-running warm air is most likely to have come from the Gulf of Mexico or the tropical portion of the Atlantic Ocean, while the underlying cold air is most frequently an air mass of Polar Canadian origin. This combination of air masses represents extremes of warmth and moisture on the one hand and of coldness and dryness on the other. The interactions of these two currents are responsible for most of the winter precipitation of the eastern part of the country.

Actually there are many modifications which are introduced into the ideal scheme shown in Figure 6. Modifications may be induced by the earth's surface features over which the front is traveling, and the vertical structure of the warm and cold air. The effect of surface friction in a retreating cold wedge (a warm front) tends to flatten the wedge in its lower sections. Thus, for some distance ahead of the warm front it is sometimes observed that the slope is almost horizontal. When this is true, a point well ahead of the front will generally be found at which the slope of the discontinuity increases comparatively rapidly. On the synoptic surface chart such a distribution may appear as another front, particularly in the precipitation field, for the steep slope causes considerable ascent of the warm air.

The vertical structure of the warm air with respect to temperature and moisture distribution is important for types of cloud and precipitation. For example, if the warm current be stable and dry, as is frequently the case with currents of Tropical Pacific air, after crossing the Rocky mountains, condensation forms may be entirely lacking, and considerable ascent of the warm air will be necessary to produce condensation. On the other hand, if the warm current is conditionally unstable, the ascent over the cold wedge may become vigorous enough to form thunderstorms. If the current be convectively unstable the lifting of layers, if carried far enough, will eventually lead to voluntary convection of the warm air, and hence shower type precipitation. In our Tropical Gulf air masses this is normally the case. One does not always find a continuous cloud layer above the warm front surface as indicated in figure 5. Rather, one often finds two or more layers of clouds separated by comparatively clear spaces. This does not invalidate the frontal idea, but merely indicates that the air is ascending the front in layers. Since the warm current is generally observed to be stratified in its temperature and moisture distribution, it is probable that, as the successive layers ascend, condensation appears at some levels sooner than at others. The cloud layer, in the event of the release of marked instability, is apt to be thick.

The structure and trajectory of the underlying cold air current, while not particularly important for the production of precipitation, is nevertheless significant for the types of clouds which are observed. Where the cold current has passed over a water surface, there will generally be found a layer of Stcu cloud, which obscures from view the upper layer of clouds associated with the warm front. The elevation of these lower clouds is rarely over 500 m, their thickness seldom appreciable, and the precipitation from them, if any, is only mist, or at most, drizzle. Occasionally, at intermediate levels within the cold air, there is found a layer of Acu clouds. These seem to be associated with surfaces of subsidence, to be discussed in a later article. In the cold air through which rain is falling, conditions are becoming increasingly favorable for the formation of clouds, for, in the first place, the moisture content is increased by evaporating rain and secondly, in the cold air adjacent to the surface of discontinuity, the lapse rate tends to steepen.

In the lowest layers of the cold air, perhaps 100 m above the surface of the earth, Frst clouds (scud) may be formed by turbulence in this almost saturated lowest layer of air.

The identification of warm fronts on the surface weather map is frequently very difficult. The wind discontinuity may not be pronounced—and to complicate matters further the transition of temperature may often appear relatively gradual. This diffuse distribution of surface elements across a warm front is explained by the small angle of slope of the discontinuity surface. At high levels it is probable that there is little mixing of the warm and cold air, but in the turbulent layer lying within about 600 m of the surface, if the discontinuity surface is not far from the ground, there must be some considerable intermingling of the warm air with the lower cold air. Thus, the temperature and moisture content of the air some distance ahead of the warm front gradually increase. A good rule to remember in this connection is that *the gradual changes which occur in the transition zone are entirely within the cold air.* Thus in the warm air current the meteorological elements, particularly temperature and moisture, should remain comparatively constant.

Upper air soundings offer the best aid to the identification of warm fronts. As pointed out in Article IV of this series, a sounding through a warm front appears on the Rossby diagram as a wedge-shaped curve with a maximum water vapor content in mid-air (fig. 5, in Article IV). The base of the warm current is generally considered as the point of the wedge on the Rossby diagram. If other soundings have been obtained within the warm air their characteristic curves should be compared with one another. This not only helps in identification, but indicates the modification which has been taking place within the air current.

VI. ELEMENTS OF FRONTAL STRUCTURE—THE COLD FRONT

Cold fronts belong to the class of discontinuities in which the warm air lying above the cold wedge is forced to rise, the energy being supplied mainly by the moving wedge of cold air. The most pronounced cold fronts are easily recognizable on the surface weather map as marked wind discontinuities, the well known wind-shift lines. On the other hand, there are many cold fronts not characterized by abrupt changes, and thus not so easily identified. The slope of the cold front surfaces of discontinuity is characteristically greater than that of the warm fronts, the values being of the order of about 1/50 in the case of the cold front compared with perhaps 1/200 for the warm front. These are merely rough averages; in any individual case the slope may be appreciably different.

The importance of fronts in weather analysis lies in the fact that they lead to the formation of vertical atmospheric motions, which in turn bring about regions of cloudiness and precipitation. It must be borne in mind, however, that no vertical motion will be present at a front which obeys the conditions of equilibrium outlined in Article V of this series. This is equivalent to saying that in order for vertical motions to arise there must be a zone of convergence. The convergence may be interpreted as the attempt of the adjacent air masses to bring about an equilibrium of frontal slope.

The typical cold front* is usually portrayed in vertical cross-section as is shown in Fig. 7. In this diagram the vertical scale is of course greatly exaggerated with respect to the horizontal. The wedge assumes the form of a squall head in its foremost section owing to the fact that the cold air is retarded by friction at the surface of the earth, while at some distance aloft, where frictional effects are small, the cold air runs ahead. The elevation of this foremost part

*According to Bergeron the cold front is of two main types: The *first kind of cold front* is an "anafront" in which there is general upsliding motion of the warm air along the front surface and formation of a rather broad post-frontal Ast-Nbst system—this type will only occur with a retarded or slowly moving front. In the "katafront", the *second kind of cold front,* the warmer air ascends only in the lower layers, the warm air above 1-4 km generally moving forwards faster than the cold wedge, sinking and thus turning the cloud system into a mainly prefrontal one of Acu-character; this kind of cold front is by far the commonest type.—*Bull. Amer. Met. Soc.*, Sept., 1937, p. 266.)—*Ed.*

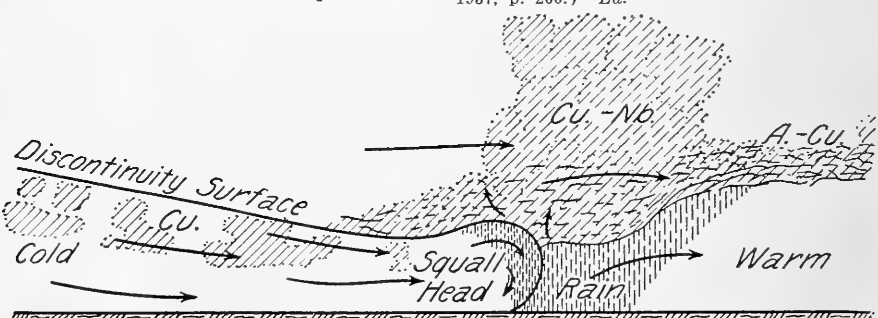

FIG. 7. IDEALIZED SCHEME OF A COLD FRONT (After Bjerknes).
(Reproduced from Fig. 91, Bull. 79, Nat. Research Council)

of the cold tongue is generally about 500 meters above the surface. The form of the advanced portion of the cold wedge depends upon the roughness of the surface over which the front is traveling, the surface cold air being the more retarded the rougher the terrain. In hilly or mountainous sections changes from the warm to the cold air are frequently less abrupt than over flat country. As the cold air tongue moves in at some upper level there is certain to be some mixing with the underlying warmer air. The existence of a cold current aloft implies a steep lapse rate, which facilitates convective stirring, and even possibly, for a brief interval of time, a superadiabatic lapse rate. It also seems logical that the cold air aloft may at times break off from the main body, much the same as the crest of a water wave. It is difficult to determine how far the cold air may run ahead of the surface boundary of the front. Perhaps twenty-five miles is an average, one hundred miles a maximum value.

The rain which occurs before the arrival of the cold air at the surface (prefrontal rain) is mainly the result of vertical lifting of the warm air by the mechanical action of the advancing wedge. The cloudiness and character of the precipitation depend upon the vertical structure of the warm air which is being displaced. If the warm current is fairly dry and stable, e.g., a current of old Polar Pacific air. the cold front may arrive accompanied only by a broken cloud deck and no precipitation. On the other hand, if the warm air is Tropical Gulf air, moisture-laden and conditionally unstable, the clouds are apt to be of the Cunb type with the characteristic frontal thunderstorms. It is evident, then, that an upper air sounding within the warm air is of importance to the forecaster.

The ceilings in advance of a well-defined cold front decrease comparatively rapidly. The foremost portion of the frontal cloud deck in the warm air may be a Ci or Cist type which rapidly lowers to Acu and Ast, and finally to Nbst. This transition may take place in a period of two hours. At times rain may be seen falling from the clouds and evaporating before it reaches the earth. The frequently observed temperature fall in advance of the cold front is sometimes associated with evaporation of rain from the clouds, the cooling extending down to the surface through turbulent mixing. When this is the case the specific humidity at the surface increases as the temperature falls. This phenomenon immediately precedes the rain.

There are times when the precipitation is entirely confined to the region ahead of the front. Here the structure of the warm air is usually of a conditionally unstable character at intermediate levels (say, 1000 to 3000 m), and the initial upward impulse some distance ahead of the surface cold front is sufficient to release and transform the potential energy into convective showers. Behind the cold front the vertical forces may be insufficient, or the structure of the warm air at higher elevations unfavorable for precipitation.

In other cases the rain zone is concentrated behind the front, i.e., within the cold air. In this case the component of wind flow normal to the front is appreciably greater in the cold air than that in the warm air. Sometimes a cold front becomes almost stationary; precipitation and cloudiness begin to spread far behind it. Pilot balloon observations well behind the front then show the warm current flowing in opposite direction to the cold air. The cold front soon retrogrades, and thus, by definition, becomes a warm front.

Normally, clearing takes place rapidly behind the cold front. The continuous cloud deck associated with the front passage soon appears to break up into Cu or Stcu clouds. Actually, however, these lower clouds are not a result of the discontinuity surface, but rather are associated with the cold air mass which is displacing the warm. The formation of these clouds is somewhat as follows. The cold air mass has come from a cold source region; therefore, it is colder than the surface over which it is traveling. Since the lower layers are continually being heated the lapse rate becomes steeper until the dry adiabat is reached. The lower particles of air then rise and cool. This process, carried far enough, leads to the formation of the Cu or Stcu clouds, and if vigorous, results in showers or snow flurries (instability showers). Soon after the front passage one may observe, through the breaks in the lower clouds, the layer of continuous cloud which lies along the discontinuity surface of the cold front. The elevation of the base of the Stcu or Cu clouds formed within the cold air remains fairly constant, but their height is somewhat less nearer the front. This is explained by the fact that the cold air near the front is more moist than the air farther back, owing to the precipitation which has fallen through the cold air and to the evaporation from the moist ground.

In traveling over different regions there are many modifications which the cold front undergoes. While there is no substitute for practical experience in interpreting the phenomena in any locality, a few more general principles can be given which may be of value.

As a cold front moves into more southerly latitudes its slope decreases in accordance with formula (1) in Article V. Thus, in the southern section of the country, it is common to find the warm air not far above the surface, even though the position of the surface boundary of the air masses is distant. The cloud deck accompanying such a front is in general horizontally extensive. Likewise, its base has a very small angle of inclination. In fact, in the southern section of the country the cloud deck associated with a slow-moving front of polar air frequently touches the ground, producing widespread fog. An airplane sounding shows a very thin layer of cold air, above which there is a large temperature inversion, the warm and moist Gulf air being aloft. The extensive fog is probably in large part due to saturation by falling rain or the mixing of the saturated, or nearly saturated Gulf air, with the cold Polar air composing the thin wedge. Precipitation falling from the warm Gulf air is, in the South, almost always in the form of rain. The cold air below the inversion, however, may possess temperatures well below freezing. In this event the rain is subcooled in its fall through the cold air, and readily freezes on striking objects. Then again it may freeze in mid-air and fall as sleet.

If the cold air passes over a warmer water surface, e.g., the Great Lakes, the rapid warming and addition of moisture generally leads to the formation of snow flurries. Conditions are most favorable for these instability flurries when the contrast between the temperature of the water and the Polar air is most pronounced, i.e., in late fall and early winter.

The flurries begin shortly after the cold air arrives and continue as long as the temperature difference obtains, and the upper wind flow carries the clouds over the particular location. It is sometimes observed that snow flurries of this sort occur when the wind at the surface is blowing off the land. In this case observations reveal the wind at levels slightly above the surface, say 500 to 1000 m, coming from a direction off the water. Evidently convection has taken place in the off-land air, which is colder than the water surface, and the resultant clouds and flurries are carried back over the land by the upper component of the circulation.

In mountainous country orographic effects upon the cold air often lead to cloudiness and precipitation. There are several factors which enter into the problem; only a brief outline can be given here. Forced vertical convection, brought about by the roughness of the surface features, produces clouds, if the lifting is sufficient. But the vertical temperature and moisture distribution in the cold air may oppose the ascent of the air. That is to say, the cold current may be dry and stable; the air will then tend to flow around obstructions rather than over them. Since vertical motion depends on the lapse rate it is logical to suppose there should be a diurnal variation in the cloudiness resulting from the above cause. During the day time, when the lowest layers are warmest, the lapse rate is steepest; on this account less force is required to produce the vertical displacement necessary to form snow flurries. At night the snow flurries either disappear or diminish in intensity.

To the lee of a mountain range the air is forced to descend, which opposes the formation of lower clouds. In these regions, therefore, rapid clearing takes place after a cold-front passage. The temperature, owing to the foehn effect, is also higher than that normally prevailing at the same level in the cold air mass.

Sometimes a cold front may degenerate into two discontinuities. In this case each discontinuity acts as a distinct front, the passage being accompanied by fairly definite changes in the meteorological elements. These *secondary fronts* are developed when the cold front is accelerating, for it can be shown that the velocity field of air flow is such that there must be a descending current of air some distance behind the main front. The descending air leads to the formation of a new discontinuity between the sinking cold air adiabatically warmed, and the fresh polar air which is not descending. The transition zone of descending air frequently appears in upper air soundings as a dry and stable layer.

While cold fronts and warm fronts have been treated independently thus far, they must not be considered as separate and disconnected phenomena. Both are necessary parts of the irregular circulation of the mid-latitudes. Both are complementary parts of the cyclone, the structure of which will be treated in the next article.

VII. ELEMENTS OF CYCLONIC STRUCTURE

In the preceding two articles of this series warm and cold fronts were discussed as independent entities. At the close of the last article it was pointed out that both are complementary parts of the cyclone of the surface weather map. The problem of the formation and maintenance of these disturbances, which are called extratropical cyclones, is exceedingly complicated and is as yet unsolved. To account for these depressions of the mid-latitudes the Norwegian school of meteorologists, under the leadership of Professors V. and J. Bjerknes, have formulated what is now generally known as the Polar Front Theory. Since the inception of this theory (during the World War) secondary modifications have been added by the Norwegians themselves and by meteorologists of other countries. In fact, at present data are being collected and studied which, within the next few years, may lead to extremely important modifications of the fundamental ideas underlying the Polar Front Theory.*

In the following article Dr. Haurwitz, an authority on the mathematical part of the Theory, interprets it for the general reader. But quite apart from their validity for a theory of the origin of cyclones (still not widely accepted), the basic conceptions of the Norwegian school have been looked upon as extremely valuable tools in the analysis of weather conditions. No impartial observer can doubt that the use of frontal ideas as a *method* of interpretation of meteorological phenomena has served to objectify and clarify for the synoptic meteorologist the description of atmospheric movements. Weather forecasting should be considered as an attempt to give first a *physical* interpretation to what *has* taken place, then, using this as a foundation, to extrapolate conditions and, what is of more importance, estimate the modifications which may occur. The method of fronts and air masses is particularly well adapted to both extrapolation and determination of modification. It thereby supplies a more quantitative basis for prognostication.

The convergence of two air masses of different properties leads to a surface of discontinuity. There are, on the earth's surface, regions where large-scale air currents of appreciably different properties generally converge. Bergeron has called these regions of *frontogenesis*. Regions of divergence, where fronts are not readily formed, and where they are generally destroyed, are called regions of *frontolysis*. For example, the region south of the Aleutian Islands may be considered as a breeding ground of fronts, because the Aleutian Low, a vast center of action formed and maintained chiefly by thermal differences between the relatively warm waters of the North Pacific and the cold snow-covered areas of Alaska and Siberia, serves to draw cold Polar Continental air into its western side and warm tropical air to its eastern side. The juxtaposition of these two air masses results in the Polar Front.

If the front formed in the Aleutian region always remained stationary under the equilibrium outlined in Article V, it would have no dynamic significance. That is to say, it would simply represent the boundary zone between the cold polar and the warm tropical air, and, since there is a balance of forces, there could be no vertical motions—hence no precipitation. It may be shown, however, that this equilibrium cannot exist, for there must be an exchange of

*Prof. Rossby has recently seriously questioned the Polar Front Theory in so far as it concerns the general circulation of the atmosphere, the flow patterns in the upper air, and the significance of fronts and air masses. That many cyclones do form from waves on a Polar Front in the lower portion of the troposphere is not denied, however. See: *Bull. Am. Met. Soc.*, 1937, pp. 201-209; *Trans. Am. Geophys. Un.*, 1937, Pt. I, pp. 130-136; and On the Rôle of Isentropic Mixing in the General Circulation of the Atmosphere, *Trans. 5th. Int. Congr. of Applied Mechanics, Cambridge, Mass., 1938*, pp. 373-8.—*Ed.*

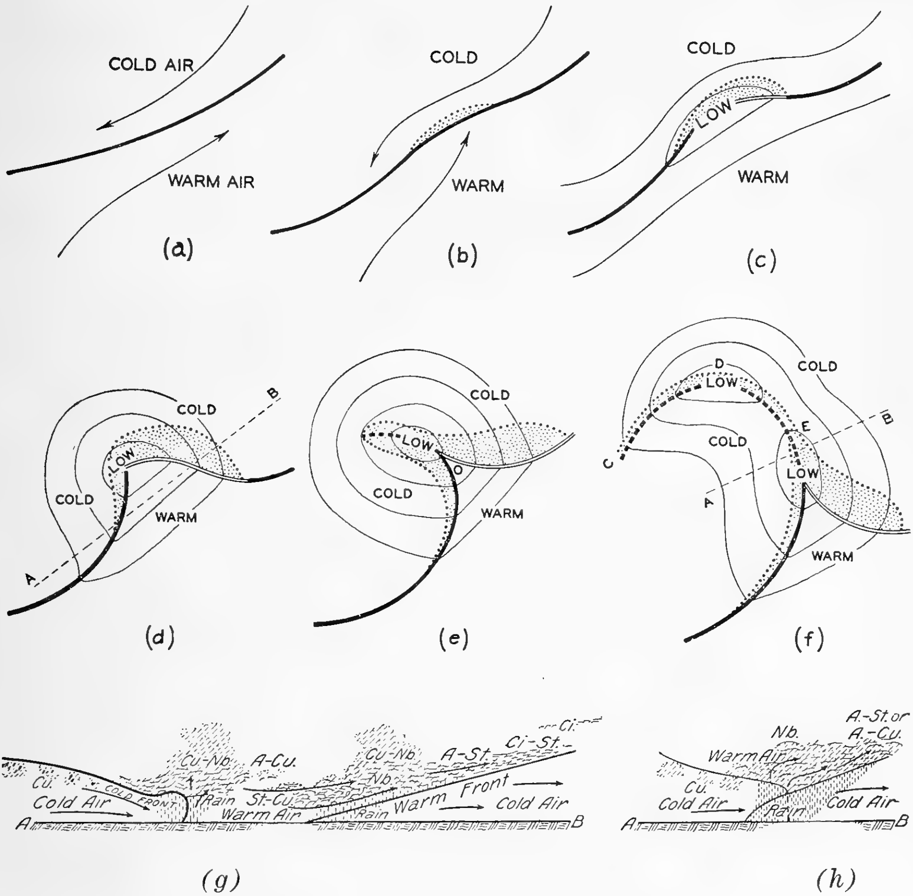

FIG. 8. FORMATION AND OCCLUSION OF A CYCLONE.

air from the polar to the tropical regions to compensate for the poleward flow which must be present in the upper levels of the atmosphere. This conclusion is based upon observations which, in the mean, show a definite poleward pressure gradient aloft. Furthermore, it can be shown that the interchange of polar and tropical air must take place sporadically and irregularly, the polar air breaking out in vast tongues and displacing the warmer air in its path. When this occurs the original low, or the primary cyclone as it is called, moves off with the system to the southeast, becoming a migratory cyclone of the mid-latitudes.

The Polar Front, which has now been pushed to the south, may extend in a general NE-SW direction, and usually slows up in its more southerly portion. We now have a slow moving discontinuity between the polar and the tropical air. This front, because it marks the boundary surface between two currents of different density and is almost stationary, is particularly favorable to the formation of wave disturbances on the front—bulges in the line of the discontinuity. The origin of these waves is to be sought in the factors which, acting together, determine the equilibrium of the frontal surface. An upset of this equilibrium at some section of the front may readily develop into a wave.*

In Fig. 8 (a) the Polar Front is shown before the beginning of a wave disturbance. Arrows represent the air flow; (b) shows the formation of a bulge in the front which is the start of a wave. The stippled area indicates where precipitation *is falling*. In this case it is the result of the vertical displacement of the tropical air over the cold underlying wedge. In the final analysis this is due to the disturbance of equilibrium. There are now two possibilities: the wave may travel along the front and remain a wave disturbance, later flattening out and dying; or, as it moves, it may increase in amplitude to such an extent that one side of it overtakes the other. This question of the stability of the wave cannot be treated in any detail here. Most of the disturbances which remain as waves along a front are those of small amplitude and short length. Longer waves of relatively large amplitude tend to close up—the cold air completely displacing the warm air from the surface.

The movement of these waves, according to V. Bjerknes, is the vector resultant of two components (Fig. 12): (1) the dynamic component of the velocity due to the displacement of the wave motion along the boundary surface and (2) the movement of the medium itself (the air in the vicinity of the front). Thus the forecasting of the motion of these waves is based upon an estimate of the mean resultant velocity of the air currents in the frontal region and an estimate of the translational velocity of the wave itself (i.e., the velocity it would have in the event the mean resultant velocity were zero). In this connection it is well to note that the component of translation due to the wave motion in the medium obeys the general physical laws of wave motion, i.e., flat, long waves travel rapidly compared to waves of large amplitude. It may also be shown that, in the northern hemisphere, the component of velocity due to the wave motion itself must be so directed that the wave moves to the right of an observer looking from the warm into the cold air. If the mean velocity of the interacting air currents also possesses this direction, it is clear that the wave will travel rapidly. On the other hand, if the resultant velocity of the air currents is directed opposite to the component due to the wave motion, the wave will move slowly, or it may even be fairly stationary. In the United States the wave component of the velocity is of more importance, for here the fronts frequently have a NE-SW direction and the cold NE current generally falls off rapidly with elevation. In the United States, then, the normal movement of wave disturbances is in a general easterly or northeasterly direction. (See Fig. 12.)

If the wave continually increases in amplitude its successive stages are represented in fig. 8 by (c), (d), (e), and (f). In these diagrams the solid heavy lines represent cold fronts; the double light lines, warm fronts; the broken heavy lines, occluded fronts (defined below); and the light solid lines, isobars. Stippled areas show where precipitation *is falling*. The right-hand section of the wave be-

*The theory of this is explained in the article by Haurwitz following this chapter.

comes a warm front, for here the warm tropical air is forced to override the cold wedge of air underlying it. The left-hand portion of the wave then becomes the cold front of the system—a wedge of cold air which is displacing the warm air. In normal cases the field of flow is such that the cold front travels faster so that it soon begins to overtake the warm front, as is illustrated at the point 0 in (e). When this occurs the process of "occlusion" is said to be taking place. Stage (d) represents the start of the occlusion. As the cold front meets the warm front the warm air lying between them is completely cut off from the surface of the earth and exists only as an entrapped body of air at upper levels. Hence it is "occluded" (closed off) from the surface. The front between the two cold air masses is called an *occluded front.* On surface charts it is represented as a line, e.g., the heavy broken line in (e) and (f).

Owing to the fact that the air currents on both sides of the cyclone have had different trajectories, the temperatures, and thus densities, of the two cold currents are not the same. As the cold front meets the warm front a new discontinuity is formed which has a shape like a wedge whose point rests on the ground, and which slopes backward or forward depending upon which current is cooler, and known as a *cold-front occlusion* and a *warm-front occlusion* respectively.

In the United States the air behind the cold front is generally colder than that ahead of the warm front, having had a shorter trajectory over relatively warm land areas. The occluded front in this case will then assume the form of a cold front in its lower portion. If, as it sometimes happens, the air ahead of the warm front is colder than the air behind the cold front, the occlusion will assume the form of another warm front in the lower layers. This latter case is illustrated in (h) which may be considered as a vertical cross-section through the line AB of (f). The intersection of the sloping front with the earth, in Fig. 8 (h), is drawn on the weather map as the "occluded front;" also, the position of the intersection of the cold and warm fronts aloft is projected on the weather map and drawn as a dashed line (in U. S. A.) known as an "upper (cold) front" (see Bergeron's model, p. 122-4). The type of weather associated with an occluded front passage will, then, depend largely upon the vertical structure. If the temperatures of the two adjacent cold currents are fairly alike, the entrapped warm air will determine the effects observed; if the warm air is still being lifted appreciably precipitation may be abundant. It is clear that the temperature and moisture distribution within the entrapped warm air is also to be considered in forecasting what is to happen. The general laws of stability are applicable. It frequently happens that an occluded front develops into a well-marked cold front. This may be explained by the fact that the occluded front represents a trough of low pressure into which there is convergence. On one side of the front, say the eastern side, the air being drawn in is becoming warmer and warmer while the air behind the front, drawn from the northwest and north, becomes colder and colder. In this manner the discontinuity at the front is intensified so that the upper trough of warm air, which was originally responsible for the front, becomes insignificant compared with the new front which has been generated in the lower layers. When this phenomenon occurs the occluded front may rightfully be called a cold front. (f) illustrates the formation of this type of discontinuity, where the occlusion has been bent south in the rear of the cyclone (in agreement with the general air flow). The section of the front from C to D gradually assumes the characteristics of a cold front, while the section from D to E becomes a warm front. When the occluded front bends around as in (f) it is called a "loop-back" or a "bent-back" occlusion. The forecaster must be on the lookout for the development of these fronts; for a time they may appear quite harmless, but later on may become important. In summer they may be the deciding factor (the "trigger action") in the formation of thunderstorms.

(g) represents a vertical cross-section through the line of AB in (d).

The following outline gives the general behavior of the meteorological elements at the surface with the passage of the different fronts. These are, of course, average conditions which are frequently observed over the *eastern* section of the United States. Individual cases may show quite different characteristics.

TABLE I

AVERAGE BEHAVIOR OF THE COMMONLY OBSERVED METEOROLOGICAL ELEMENTS WITH FRONT PASSAGES OVER THE EASTERN UNITED STATES IN WINTER.

Element (at surface)	Type of Front Passage: Warm Front	Cold Front	(Cold) Occluded Front
Pressure	Falling, then steady or falling less rapidly	Falling then slightly rising, then abrupt rapid rise	Falling slowly, then rising slowly
Temperature	Steady, then rising at an increasing rate, then steady	Rising slowly, then falling slightly, then dropping abruptly at a rapid rate	Not much change, falling slightly after front
Relative Humidity	Fairly constant, then increasing to nearly saturation, then falling	Increasing gradually to almost saturation, then falling	Gradually increasing to almost saturation, then falling off
Specific Humidity	Steady, then increasing at an increasing rate, then constant	Fairly constant, slightly increasing, then falling abruptly and rapidly	Slight increase then general decrease
Clouds	Ci to Cist to Ast to Nbst to Frst	Cist rapidly changing to Cunb, then to Frcu	Ast to Nbst to Cu
Precipitation	None, then gradual steady rain, then rapidly falling off to none	None, then heavy but brief shower, perhaps thundershower, clearing rapidly	Gradual, fairly steady, but not long lasting
Visibility	Fair, becoming poor, and remaining poor	Poor, then rapidly becoming good to very good	Decreasing to poor, then increasing to fair or good
Wind	SE or S to SW	S or SW to NW	W to NW

The physical interpretation of these changes in terms of what has been said about fronts and air masses is left to the reader.

The Norwegian Wave-Theory of Cyclones[1]

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In recent years air mass analysis has been widely accepted by meteorologists on this continent as a basis for the study and forecasting of the weather. One of the fundamental assumptions of this method is the theory that cyclones form as waves at surfaces of discontinuity between air masses of different origin and history which, consequently, have different physical properties, e. g., temperature and humidity.

The practical side of air mass analysis is well represented in this booklet by Mr. Namias and Prof. Willett. The theoretical questions, whether waves actually can occur at such surfaces of discontinuity and whether these waves have any resemblance to the nascent cyclones observed in the atmosphere, have only been touched slightly in their articles. They are, naturally, of secondary interest to the practical meteorologist.

The Problem

Perhaps it might be argued that a mathematical study of the possibility of cyclonic waves is just a fruitless pastime without any practical value. Are not cyclones in their earliest wave-like stages found on almost every weather map? However, before the Norwegian school of meteorology conceived their idea of the origin of cyclones as waves at frontal surfaces, no such phenomena were noticed. If something appeared on the map which today would be interpreted as a frontal wave it certainly was formerly never recognized as such. And even now the wave theory of cyclones is not generally accepted. Thus there can be no doubt that the theory of the origin of cyclones as frontal waves can be strengthened and accepted only if it is proved that such atmospheric oscillations are actually possible and do occur more or less generally in the atmosphere.

Furthermore, the quantitative relations between length and velocity of the cyclonic wave and temperature and wind discontinuity at frontal surfaces can only be obtained by the mathematical theory. The verification of such formulae by observations would represent the real check on the theory. Quantitative relations are further necessary since there exists sometimes a tendency to interpret weather situations without regard to the obvious numerical impossibility of the proposed explanation.

Thus the wave theory of cyclones presents a definite hydrodynamical problem to the theoretical meteorologist. He has to show that at frontal surfaces such as we observe in the atmosphere, waves can originate which have the characteristic qualities of the cyclonic frontal waves which are found on the maps; they must possess a similar wave-length and velocity of propagation, and the motion of the air particles in the computed wave must agree with the air motion in the observed nascent cyclones. Another very important point is that these waves must be unstable, i. e., their amplitude, at first very small, must increase with time until the cyclone loses its wave character and becomes a vortex. The later vortical stages of the life his-

[1]This article first appeared in the Bulletin A.M.S., June-July, 1937.

*Research Associate, Blue Hill Observatory.

tory of a cyclone (when the warm sector occludes) have not yet been attacked theoretically owing to the extreme difficulty of the mathematical treatment. But the nascent waveform has already been dealt with very thoroughly from a mathematical viewpoint. While it must be admitted that much still remains to be done on the mathematical analysis of the young cyclone, as will become apparent later, it has been proved, mainly by H. Solberg[2]), that waves postulated in the Norwegian (wave) theory of cyclone formation can and must exist in our atmosphere.

In the following the results of the mathematical theory will be represented in a descriptive form. Readers who desire to study the complete theory are referred to "Physikalische Hydrodynamik" by V. Bjerknes and his collaborators[2]). Here the various types of waves will be described which originate in a liquid or a gas due to different causes, and it will be shown how these different causes act together in the atmosphere to give origin to the cyclonic waves. In this way we follow somewhat the same line of attack as used by the theoretical meteorologists. Owing to the complexity of the problem it has been necessary first to disregard some of the influences acting upon the waves in order to study somewhat simpler conditions.

Gravity Waves

A well known type are the waves at a water surface, or, to be more accurate, the waves at the boundary between water and air. For wave motions of this type, the *gravity* of the earth is the controlling force (disregarding the small ripples which are subjected to capillary forces). To understand better the physical process in a *gravitational wave* consider a mass of water which is lifted upwards from the original position of equilibrium by a disturbance. This quantity of water has thus acquired a certain potential energy. While falling back to its original place its potential energy is transformed into kinetic energy of motion. The velocity increases until the water passes through the equilibrium position. However, it does not come to rest but owing to the acquired velocity continues to move downwards until the kinetic energy is transformed again into potential energy. Then the mass element reverses its direction of motion and ascends. In this way a wave motion is set up which is damped out gradually only by friction. This description applies to standing waves since it takes only vertical motion into account. For progressive waves it would have to be modified slightly. Moreover, it is not quite satisfactory to single out one mass element and neglect the motion of the surrounding water masses. These defects not withstanding, it shows that there is always a transformation from kinetic to potential energy and back in a gravitational wave, analogous to the oscillations of a pendulum.

Similarly, the internal waves at the boundary between two fluid layers of different density are gravitational waves. Waves of the gravitational type are stable as long as the original density distribution is stable. Unstable stratification is rarely found in the earth's atmosphere except near the ground, for obvious reasons. Therefore, gravity tends to generate only stable waves.

[2]Bjerknes, V., Bjerknes, J., Solberg, H., and Bergeron, T.: Physikalische Hydrodynamik Berlin, 1933, pp. 565-621; Hydrodynamique Physique. Paris, 1934, 3 vols., Chapt. 14, (vol. 2).

Compressibility

The compressibility of water is comparatively small so that it can be neglected in most cases. Atmospheric air, on the other hand, has a high degree of compressibility. Under the influence of compressibility alone sound waves are obtained. The influence of compressibility on atmospheric wave motions does not give rise to unstable waves as long as the lapse rate of temperature is below the adiabatic, or for saturated air the moist adiabatic. The considerations which lead to this conclusion are well known, and were given elsewhere in this booklet[3]).

Shearing Waves

If a wind discontinuity is present in an otherwise homogeneous fluid, waves are possible at this surface. These waves are always unstable in contrast to the wave types we have mentioned so far, which were only unstable if the stratification of the fluid or the gas was unstable to begin with. The computation shows that the amplitude increases more rapidly with time, or in other words, the wave is more unstable the smaller the wave length. Since the existence of this type of wave is due to the wind discontinuity or to the shear of the wind, it will be called a *shearing wave* here, and we shall speak of *shearing instability.* In the atmosphere, hardly ever do we observe a shearing discontinuity which is not associated with a temperature inversion and therefore a density discontinuity.* When this is the case, waves at the surface of discontinuity must be of a mixed type, since the stabilizing gravitational effect and the unstabilizing shearing effect act at the same time. For small waves the unstabilizing effect of the wind shear overcompensates the stabilizing effect of stratification and the waves are unstable. As we pass on to longer waves we come into a region where the stable stratification is more effectual than the shearing instability. So now we are in the region of wave lengths for stable waves. The limit between shorter unstable waves and longer stable waves lies at a wave length of at the most a few km under atmospheric conditions. It varies of course with the order of the wind and temperature discontinuity. It will be seen from this discussion that the unstable waves of about 1000 km length required by the cyclone theory are not obtained when only the influence of gravity, compressibility and wind shear are considered.

Inertia Waves

In order to find waves which resemble the nascent cyclones of the weather map the effect of the earth's rotation has to be taken into account. To understand fully the importance of the earth's rotation upon atmospheric wave motions we start out with a simple problem which at first seems quite devoid of any meteorological application. Owing to its fundamental significance it will be necessary to deal with this case somewhat in detail.[4]) If a hollow cylinder is partly filled with water and completely closed, then set in rotation around a central vertical axis, the surface of the fluid which at rest was horizontal will become parabolic. When the rotation is sufficiently fast, the water is pressed completely against the cylinder walls so that the liquid has a practically cylindri-

*Even if no temperature inversion is present there is still a certain degree of stability as long as the vertical temperature gradient is less than the adiabatic.

[3]Namias, this booklet, pp. 4-8.

[4]Bjerknes, V. and Solberg, H.: Zellulare Trägheitswellen und Turbulenz, *Avh. Norske Vid. Akad.*, Bd. I, Math-Nat. Kl., no. 7, 1929, pp. 1-16.

cal shape. The gravitational force is so much smaller than the centrifugal one that its effect is unnoticeable when the rotation is sufficiently fast. Instead of the horizontal surfaces of liquids as ordinarily observed in nature under the vertical action of gravitation, the fluid in the rapidly rotating cylinder has a practically vertical surface due to the horizontal action of the centrifugal force. If this vertical surface is subjected to a small disturbance a wave motion will originate. This is quite analogous to the case of gravitational waves on a horizontal surface of water, but in that case the energy of the wave motion was gravitational. In the present case of a rotating cylinder the centrifugal force replaces the gravitational force. Since the centrifugal force is due to the inertia of the mass these are called *inertia waves* by V. Bjerknes and H. Solberg.[4])

The existence of inertia waves in a rotating fluid can also be demonstrated by an elementary computation. If a homogeneous incompressible fluid like water is enclosed between two rigid walls which are infinite in horizontal direction it can easily be shown that no wave motion is possible as long as the fluid system does not rotate. As soon as the rotation is taken into account, as is necessary for the large scale atmospheric motions on the rotating earth, it is found that now a wave motion is possible with a period longer than half a pendulum day. (A pendulum day is equal to 24 sidereal hours divided by the sine of the geographic latitude, this being the time required for the swinging plane of a pendulum on the rotating earth to return to its initial position.) Only the angular velocity of the earth's rotation and not the acceleration of gravity appears in the relation between velocity and length of this type of waves, indicating that they are inertia waves.

Some remarks on stability and instability of inertia waves are necessary in order to see their significance for the wave theory of cyclones. To choose the simplest case, which shows the principle clearly enough, it may be assumed that a fluid mass rotates around a vertical axis with an angular velocity q which is constant for the whole fluid. If the fluid is enclosed between rigid horizontal boundaries and situated at either pole of the earth we have just the case considered in the previous paragraph. The constant angular velocity of the fluid is equal to the angular velocity of the earth's rotation. An observer on the earth does not observe this "absolute" velocity since he takes part in the rotation of the earth.

The "absolute" velocity v at the distance r from the center is

$$v = q.r$$

From the principles of mechanics it is known that the angular momentum $vr = q.r^2$ of an individual particle remains constant. Thus, if a particle is pushed away from the axis, say from the distance r to $r + s$, then the constancy of the angular momentum requires that this particle in its new position must have a smaller angular velocity q' which is given by

$$qr^2 = q' \, (r + s)^2$$

while the angular velocity of the surrounding fluid masses at the distance $r + s$ is q as before.

The centrifugal force acting on the displaced particle is

$$q'^2 \, (r+s) = q^2 \frac{r^4}{(r+s)^3} = q^2 r \quad (1-3\frac{s}{r})$$

approximately,

while the centrifugal force in the surrounding fluid at the same distance is greater, namely,

$$q^2 (r + s)$$

Thus the displaced particle has a deficit of centrifugal force which drives it back to its original position where it is in equilibrium with its surroundings. Similarly when a particle is displaced towards the axis it obtains a surplus of centrifugal force as compared to the surrounding fluid and will again move towards its initial position. Thus the origin of inertia waves is easily understood. If a particle is displaced outwards from its equilibrium position it is driven back to this position. But while moving back to the equilibrium position it acquires kinetic energy so that it approaches the axis of rotation more closely than before the disturbance until the surplus of centrifugal force which the particle gains in approaching the axis overcomes the kinetic energy and reverses the direction of motion again. The process is completely analogous to the wave motion in the atmosphere in stable equilibrium, except that there the deficit or surplus in weight as compared to the surrounding air plays the rôle of the centrifugal force.

Earth's Rotation

It is obvious from the previous considerations that inertia waves in a fluid rotating with constant angular velocity are stable. If our atmosphere were at rest with respect to the surface of the earth, it would appear to an extra-terrestrial observer to rotate with constant angular velocity around the earth's axis. In reality, owing to the different winds in the atmosphere, or, in other words, to atmospheric motions relative to the earth, the atmosphere does not have strictly speaking a constant angular velocity. However, closer inspection shows that the wind distribution in the atmosphere is such that inertia waves of the dimensions of cyclonic waves are always stable. With Bjerknes and Godske[5]) we may call this the *"dynamic stability"* due to the earth's rotation, while the stability of the gravitational waves in the earth's atmosphere might be termed *"static stability"*.

Another important effect of the earth's rotation is the following. In purely gravitational waves the motion of the fluid particles takes place in a vertical plane, for the direction of the gravitational force is vertical. The deflecting force of the earth's rotation, on the other hand, acts perpendicular to the earth's axis. Thus it acts purely vertically only at the equator. At the pole it is directed horizontally, but at all other latitudes it is inclined to the horizon, so that we can speak of a horizontal and a vertical component. The vertical component generally can be neglected in meteorology since its direction coincides with the far greater acceleration of gravity. The horizontal component of the deflecting force of the earth's rotation causes an inclination of the plane of motion with respect to the vertical. The inclination is larger the greater the motion; thus in the long cyclonic waves the motion is predominantly horizontal, while in the small billow clouds it is vertical.

Cyclonic Waves

Now that the different factors which cause and influence wave motion have been considered separately it can more easily be understood how they act together to produce the cyclonic waves.

It was stated that according to the Norwegian wave theory we have frontal surfaces between air masses of different density (temperature, moisture content). Due to the different densities on opposite sides of the

[5]Bjerknes, J.; and Godske, C. L.: On the theory of cyclone formation at extra-tropical fronts, *Astrophysica Norvegica*, Vol. I. no. 6. 1936, pp. 218-219.

frontal surface the pressure gradients and consequently the winds are likewise different; these are observed facts. The observations also indicate strongly that the cyclones originate from unstable waves on such frontal surfaces. The possibility of unstable waves with the characteristic dimensions and motions found in nascent cyclones, however, has to be proved. The theory shows that very small waves are unstable up to a length of a few 100m, or, if the vertical stratification is not very stable, a few km. Exact figures can not, of course, be given as long as the density and wind distribution are unknown. The limit between stable and unstable waves depends on the lapse rates of temperature in both air masses, and on the temperature and wind discontinuities at the frontal surface. But it can be stated that for sufficiently small waves the shearing instability (wind discontinuity) is more effective than the gravitational stability. In other words, very short waves are unstable and have more the character of shearing waves than of gravitational waves. For the theory of the cyclone these waves are evidently far too short. When the waves are longer the stabilizing influence of gravitation becomes more pronounced so that beyond a certain critical wave length the waves will be stable, because the effect of shearing, which tends to produce instability, decreases with increasing wave length and is over-compensated by the stabilizing effect of gravitation. The gravitational-wave character is now predominant. Billow clouds are waves of this type.

The wave length of billow clouds is the greater, the smaller the temperature discontinuity. From this it has been concluded that the wave theory of cyclones breaks down since wave lengths of the order of 1000 km could only exist for infinitely small temperature discontinuities and impossibly large wind discontinuities, according to the formula for billow clouds. The fallacy of this objection can now be seen immediately. In the investigation of the billow waves the influence of the earth's rotation is neglected and can be neglected owing to the small dimensions. But waves of cyclonic dimensions are greatly modified by the rotation of the earth around its axis.

Thus, the wave motion becomes with increasing wave length more and more inclined to the vertical under the influence of the deflecting force of the earth's rotation. This leads to the formation of unstable waves in the following way. The stable character of gravitational waves in the earth's atmosphere depends on the difference between the weight of the oscillating particle and that of its surroundings and thus on the vertical component of the oscillation. Consequently, as the waves become longer and the wave motion more horizontal the stabilizing effect decreases, since the vertical component of motion decreases. The computation shows that waves whose length is of the order of 1000 km are unstable. At such wave-lengths the shearing instability is greater than the gravitational stability (which is small) owing to the almost horizontal motion of the particles. But with still longer waves the shearing instability becomes smaller than the dynamic stability caused by the earth's rotation and, therefore, such waves become stable again.

It is obvious that the only waves which may be regarded as cyclonic are the unstable ones whose length is of the order of 1000 km. They are of a mixed type because, in addition to shearing instability, gravitation and inertia act upon them.

The mathematical analysis also shows that these waves have velocities of the order of magnitude observed in nascent cyclones, that the velocities are generally directed eastwards, and that the motion of the air is of the type found on the weather chart. Therefore, we can say that the theory shows that the formation of a cyclone from waves not only is *possible* but *must* occur in the atmosphere because the waves are frequently unstable and therefore form spontaneously. Of course, stable waves also occur; if they are much shorter than cyclonic waves they are observed as billow clouds, ceiling fluctuations, and microbarometric oscillations, and if longer than cyclonic waves (and thus stable again owing to the stabilizing influence of inertia) they appear as flat frontal waves like young cyclones, but never develop into mature cyclones. (See Fig. 9.)

The theoretical investigation of cyclonic waves shows further that the velocity of propagation consists of two terms. Their physical significance is understood most easily by considering a wave on the surface of a river. The propagation of such a wave is partly due to the bodily transport of the oscillating water masses by the flow of the river, the "convective" term, and partly due to the motion of the wave relative to the water, the "dynamic" term. Similarly, the first term in the expression for the wave velocity at a surface of discontinuity is due to the mean motion of both air masses. It simply indicates the fact that the wave is transported passively due to the undisturbed motion of both layers. This part of the total velocity of propagation is therefore called the "convective" term. The second term, which is due to the dynamical processes of the wave motion, is referred to as the dynamic term.

Remaining Problems

While the wave theory is sufficiently far advanced for one to say with certainty that cyclonic waves occur in the atmosphere, much remains to be done yet. The investigations so far have been dealing with wave motions at the boundary between isothermal air masses, because isothermy gives equations which are easier to handle than equations which take the usual linear temperature gradient into account. But isothermal lapse rate implies, obviously, a larger stability of atmospheric stratification than is ordinarily found, so the numerical results will have to be modified. Notable advances in this direction have recently been mode by H. Solberg.[6]) Furthermore, note that a sharp discontinuity of wind and temperature is assumed in the theory while in reality a narrow transitional zone exists in which the elements change rapidly but continuously. It has been shown, however, that the wave is practically the same whether there is a sharp discontinuity or a transitional zone, provided that the thickness of the transitional zone is small compared with the wave length[7]). This condition is always fulfilled for cyclonic waves.

A certain deficiency of the wave theory of cyclones in its present state, which we tactfully have not mentioned so far, is the assumption that the lower cold layer and the upper

[6]H. Solberg: Schwingungen und Wellenbewegungen in einer Atmosphäre mit nach oben abnehmender Temperatur, *Astrophysica Norvegica*, vol. 2, no. 2, 1936, pp. 123-172.

[7]Haurwitz, B.: Zur Theorie der Wellenbewegungen in Luft und Wasser, *Veröffentl. d. Geophysikal. Inst. Univ. Leipzig, Spezialarb., Ser. 2,* Vol. V, no. 1, 1931, pp. 52-53, 73-74.

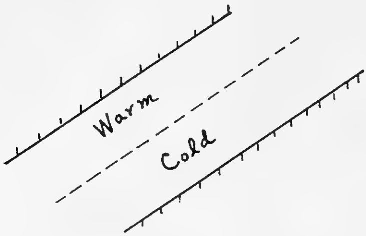


Fig. 1

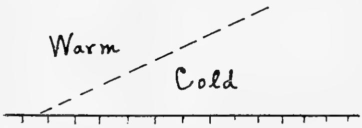


Fig. 2

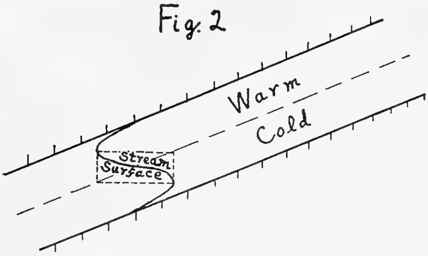


Fig. 3

Fig. 1. (9) Vertical Cross Section Through a System of Two Air Masses with Parallel Boundaries Separated by a Surface of Discontinuity, as Considered by the Theory. The Boundaries and the Surface of Discontinuity are Inclined Against the Surface of the Earth.

Fig. 2. (10) Vertical Cross Section Through the Actual Position of Warm and Cold Air Masses Relative to the Surface of the Earth.

Fig. 3. (11) Same as Fig. 1 with a Stream Surface Which Runs Partly Horizontal Over This Approximately Horizontal Area (Indicated by the Rectangle) the Stream Surface May be Regarded as the Surface of the Earth.

warm layer have boundaries parallel to the frontal surface (Fig. 9). In reality, the frontal surface intersects the surface of the earth so that the cold mass has the form of a wedge (Fig. 10). Solberg[8]) has given some preliminary results but the problem is so extremely difficult that until now a somewhat round about approach has been used. Waves of the cyclonic type in the layers parallel to the frontal surface and therefore inclined to the horizontal according to Margules' formula[9]), have been studied. It was found that some of the "stream surfaces" along which the motion of the air particles takes place are practically horizontal planes for an extent of about 2000 km or more (Fig. 11). Now a stream surface can be assumed to be rigid, since according to the definition of a stream surface the motion is parallel to it. If we let the stream surface in Fig. 11 become solid and regard it as the surface of the earth a very close analogy to the atmosphere is obtained in the region near the intersection between stream surface and surface of discontinuity. The cold air lies like a wedge under the warm air and the wave motion near the surface is almost parallel to the solid horizontal stream surface, which may be identified with the surface of the earth. Farther away from the frontal surface the agreement with reality is of course less satisfactory, since the curvature of the solidified stream surface will be stronger. But the intensity of the wave decreases with the distance from the surface of discontinuity. The field of motion is dynamically most important at the front, and loses its significance comparatively rapidly in lateral and vertical directions. Therefore the method of assuming solidification of a horizontal stream surface is more satisfactory than might appear at first.

[8]Solberg, H.: Integrationen der atmosphärischen Storungsgleichungen (I), *Geofys. Publ.*, vol. V, no. 9, 1928, pp. 104-120.
[9]Namias, this booklet, p. 25.

Nevertheless it will be necessary to solve directly the problem of a wave motion at a frontal surface inclined to the surface of the earth, especially in order to obtain reliable quantitative criteria to decide whether an observed wave is stable or unstable.

Finally, cyclonic waves have thus far been investigated only on a rotating plane. Considering their dimension, however, it is to be expected that the curvature of the earth has a certain influence. In this respect also much remains yet to be done. But the possibility of cyclonic waves in our atmosphere can be regarded as proven in spite of these gaps in the theory, and it can safely be said that objections to the wave theory of cyclones result from insufficient knowledge of the theoretical investigations.

Frontal Waves

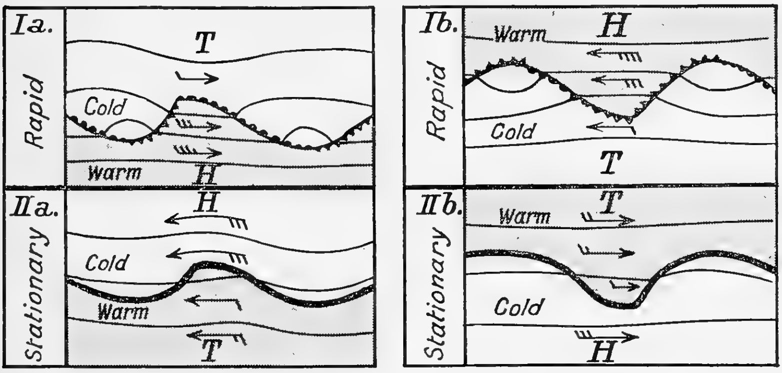


FIG. 12. FOUR TYPES OF FRONTAL WAVES.—It follows from the wave-theory (see article by Haurwitz, above) that the propagation of a frontal wave can be divided into a *dynamical part,* which always is directed eastward with a normal temperature distribution (colder to the north) and westward with a reversed one, and an *advective part,* being approximately equal to the mean of the velocities of the two air masses surrounding the front. Hence one gets 4 types of frontal waves: (1) *rapid waves,* forming in a Wly current with colder air polewards (*Ia*), or in a general Ely current with an inverted temperature distribution (*Ib*).; and (2) *quasi-stationary waves,* forming in a general Ely current with normal temperature distribution (*IIa*), or in a general Wly current with inverted temperature distribution (*IIb*).These are all frequently observed on synoptic maps.—From: Bergeron, "Physics of Fronts," *Bull. Amer. Met. Soc.,* Sept., 1937, pp. 269-70.

Sources of Energy for Extratropical Cyclones

Dr. Bergeron of the "Norwegian School" recently stated. . . . . "that the process of cyclogenesis derives its energy mainly from three sources:

(1) The kinetic energy of the preexisting air currents on both sides of the cyclogenetic front ("current energy") — insufficient, however, alone to explain the display of energy in intense cyclones.

(2) The potential energy of the hozizontal distribution of mass ("frontal energy"), which is transformed into kinetic energy according to the Margules' principles of energy—probably the main source of energy.

(3) The potential energy of a moist-labile [i.e., unstable for saturated air] stratification,

"Labilitätsenergie" in the terminology of Refsdal.†

It is shown that the lability energy† though smaller than the current and frontal energy, probably plays a great role in cyclogenesis, because it may get stored up a long time beforehand over a rather great area, and only released at the most favourable moment and in the very center of the cyclogenesis, without much frictional loss of energy.

The author has tried to form a more concrete idea of the *mechanism of cyclogenesis*, i.e., of the fall of pressure around the "warm tongue" of a frontal wave of increasing amplitude.

This mechanism should consist in a sinking of the centre of gravity of the whole system under consideration, conditioned by that ascent and convergence of warm air at the "warm tongue" of a frontal wave of increasing amplitude. This convergence implies in its turn an increase of cyclonic circulation within this area, which again involves a fall of pressure in the middle of it, compensated by a slighter general rise of pressure in the outskirts of the disturbance." (*Bull. Amer. Met. Soc.*, Sept. 1937, p. 269.)

†According to Refsdal ("Der feuchtlabile Niederschlag", *Geofy. Publ.*, Vol. 5, No. 12, 1930, p. 9) the *"lability energy"* of an air mass is defined as the maximum amount of energy which could be set free by an overturning or an upward motion of this air mass. The "lability energy per unit mass" between two heights is the algebraic sum of the work produced when a particle of unit mass is raised from the lower to the higher level (the compensating downward motion of the surrounding air being spread out over an infinitely great area). This is conditional instability, including the instability realizable by condensation of water vapor. Compare with discussion of energy in Articles on the Tephigram and Thunderstorm.

A Note on Dynamic Anticyclones and Cyclones

In the discussion of American air masses it is pointed out that cold Pc air is characteristically shallow. In fact the majority of moving anticyclones on American weather maps are of the shallow so-called cold (or polar) type, having warmer Pm or Tm air (with often even low pressure and cyclonic winds) aloft.* Wexler† has shown why the radiational cooling in the polar regions does not succeed in building up a very deep layer of Pc air before it is released to lower latitudes; the upper warm layers that may move south bodily with the Pc are thus Pm or Tm air still unmodified to Pc. Subsidence‡ usually greatly accentuates the *upper* warmness of these cold-type highs as they move southward. Figs 3-6 herewith, from Haurwitz and Noble, show a case of a cold surface high replaced by low pressure aloft. Note, however, that this is in part due to the backward slope in the free air of the axes of the lows, since the lowest pressure is always at the cold front at any level and the front itself slopes backward.

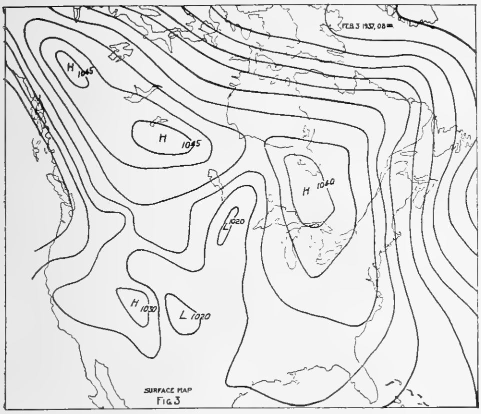
However, there are also so-called warm anticyclones which in most cases seem to have developed from cold ones after the latter have slowed down and suffered heating from below in middle latitudes. Since these highs are warmer than cyclones from the surface up to high levels (8 km or more) the maintainance of their anticyclonic winds and the high surface pressure cannot be explained thermally. It must be due either to some dynamic process which piles up air in the upper or middle troposphere or to a northward advection of cold (tropical) air in the stratosphere over the anticyclone, sufficient to overcompensate the lowering of pressure from warming in the troposphere. The warm highs are common in western Europe, where the cold-stratosphere advection theory for their maintenance has been generally advocated in recent years. Indeed, some European cases** have been studied (from soundings) where *apparently* a tremendous rise in stratospheric pressure caused a marked day to day rise in the surface pressure of a cold anticyclone while it was becoming warm (a common phenomenon); but it has not yet been proven that this was more than a coincidence rather than some necessary consequence of the effect of tropospheric pressure changes in inducing stratospheric advection or vice versa. In an American case studied by

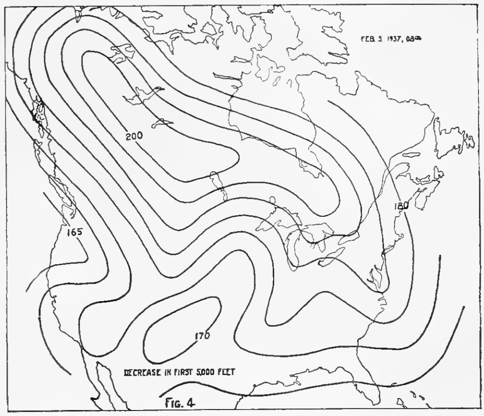
*Haurwitz and Noble, *Bull. Amer. Met. Soc.*, March, 1938, pp. 107-111.—A good example; J. Namias, Structure of a Wedge of Continental Polar Air, *M. I. T. Met. Course, Prof. Notes No.* 6, 1934.

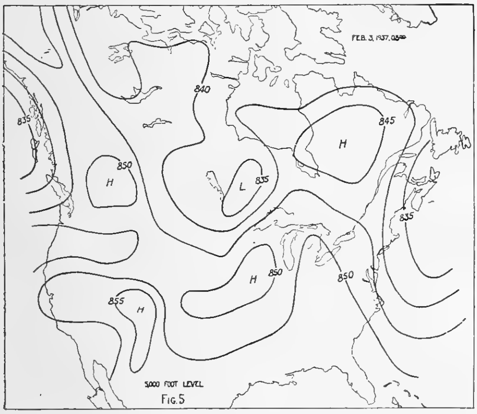
†H. Wexler, *Mon. Wea. Rev.*, April 1936, p. 122-135. and June 1937, pp. 229-236.

‡J. Namias, Subsidence in the Atmosphere *Harvard Met. Studies* No. 2, 1934.

** H. Thomas, *Sitzber. Preuss, Akad. Wiss., Phys. Math. Kl.* vol. 17, 1934; Khanewsky. *Met. Zeit.*, 1929, p. 81; Runge, *Met. Zeit.*, 1932, p. 131; Schmiedel, *Veröff. Geophys. Inst. Univ. Leipzig*, vol. 9, no. 1.







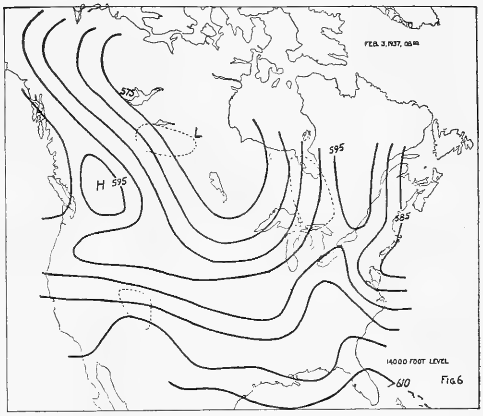


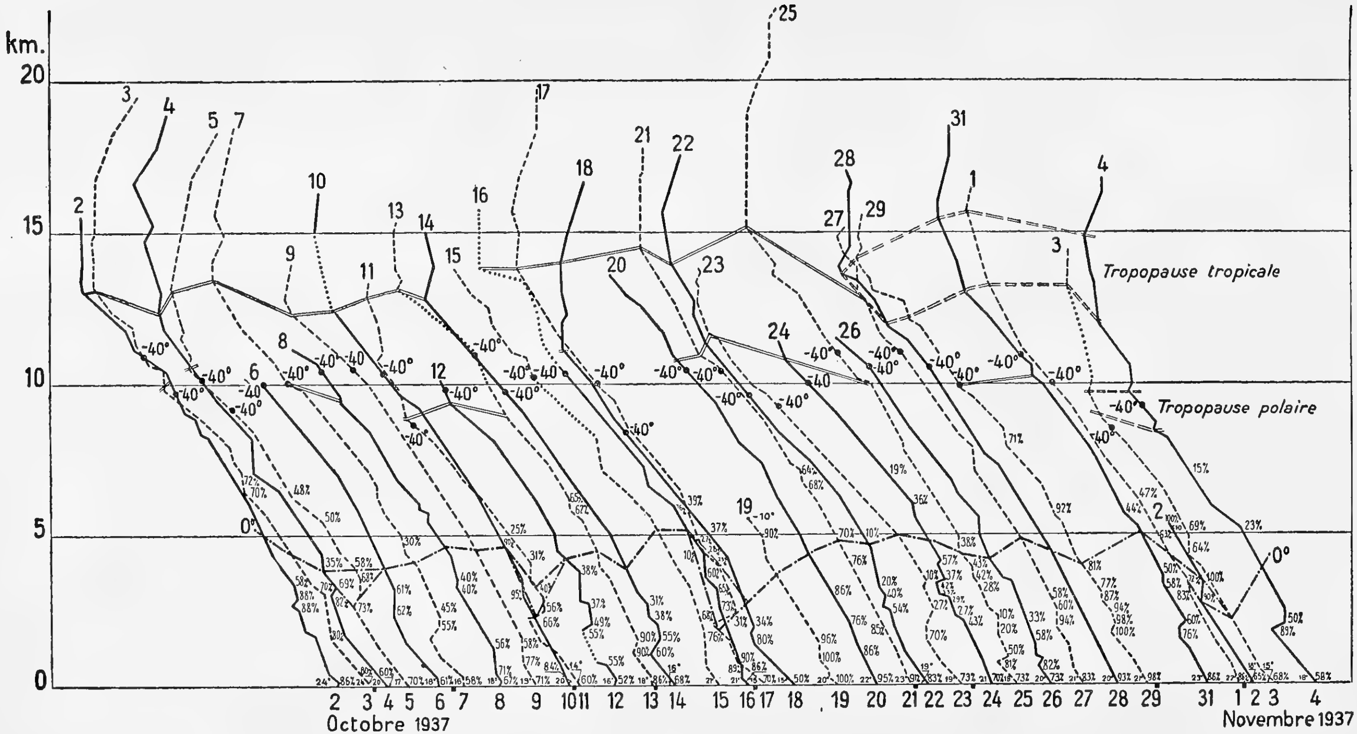
FIG. 3. PRESSURE DISTRIBUTION AT SEA LEVEL (mb), FEBRUARY 3, 1937, 8 A.M.
FIG. 5. PRESSURE DISTRIBUTION AT 5,000 FEET, FEBRUARY 3, 1937, 8 A.M.

FIG. 4. PRESSURE DECREASE IN FIRST 5,000 FEET, FEBRUARY 3, 1937, 8 A.M.
FIG. 6. PRESSURE DISTRIBUTION AT 14,000 FEET, FEBRUARY 3, 1937, 8 A.M. (The dotted circles show the position of the surface Highs)

Simmers* using isentropic analysis, a warm upper-level anticyclone over Texas and a cold one moving down from Canada amalgamated. The warm anticyclone built up from middle to high levels through a process of isentropic mixing which transferred air across isobars ("banking effect" described by Namias in his chapter on Isentropic Analysis). The resulting convergence caused the surface pressure to continue to rise, although the cold anticyclone was being dissipated by surface heating. This was in May and there was no evidence the warm anticyclone grew from the cold one, but rather their paths just happened to intersect. Probably there are various possible types of anticyclogenesis.

Cyclones also occur in warm and cold types. The usual open warm-sector or freshly-occluded low contains a warm core, but frequently an occlusion develops into a deep slowly-moving vortex without fronts but cloudy and with precipitation; soundings will show the core to be *colder* than the surroundings up to high levels, even to the stratosphere, which may be sucked down and warmed, thus helping to maintain the surface low. These cold cyclones occur particularly in the regions where cyclones usually have reached the occluded stage, in the higher latitudes of the

*R. G. Simmers, *in:* Fluid Mechanics Applied to the Study of Atmospheric Circulation, Part I, *Papers in Phys. Oceanogr. and Met.*, vol. 8. 1938.



Le signe ▮ *indique le passage d'un front froid* (début d'une invasion polaire).

FIG. 1. Radiometeorograph Soundings in the middle North Atlantic between Oct. 2 and Nov. 4, 1937; showing temperature-height curves, relative humidities, 0° C and –40° C isotherms, tropopauses, and cold fronts (black squares at base). The ship was in the square, 38°–43° W 38°–44° N. during this period. These data show the tendency for low (polar) tropopause with warm (Arctic) stratosphere air to blow in over a tropospheric cyclone, while high (tropical) tropopause with cold (Equatorial) stratosphere air appear over a tropospheric anticyclone. Since the tropopause slopes S to N, this can be due to large waves set up at the tropopause levels and propagated eastward. Whether these waves are merely induced by the effect of passing highs and lows in the troposphere, or at stages (as over warm anticyclones) they also may tend to "steer" the surface pressure waves has been a question of great controversy in Europe; both effects probably play a rôle. Also the tropopause is often sucked down adiabatically by deep cyclones (Palmén). These soundings were made by the French floating meteorological station on the S.S. *Carimaré*. (*Note*: the temp. scale is shifted 10° C to the right from one sounding to next.)

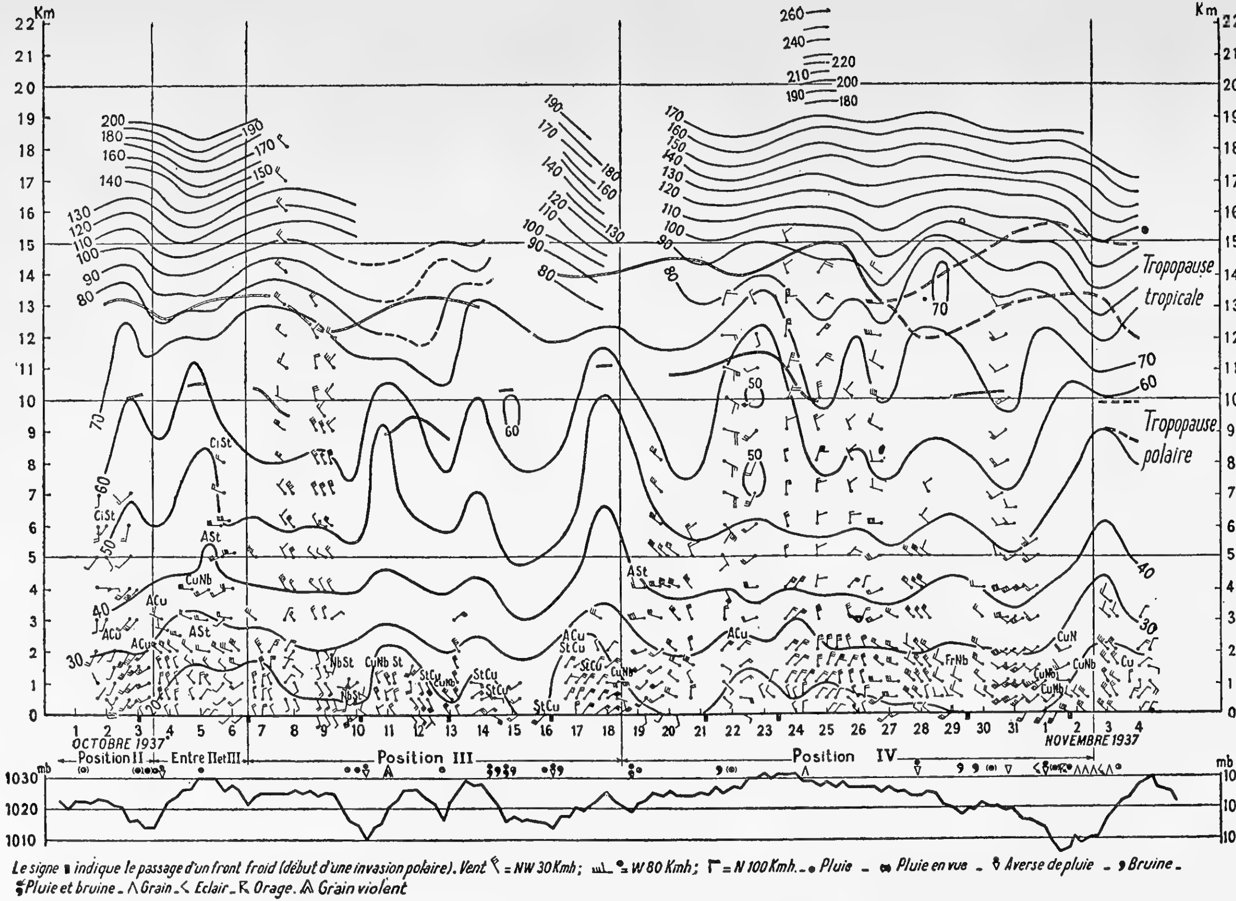


FIG. 2. Same soundings as in Fig 1, showing isotherms of potential temperature, tropopauses, upper winds, clouds, and weather. (Weather in international symbols; pluie=rain; grain=squall; eclair=lightning; orage =thunderstorm; averse=shower; bruine=drizzle). Surface pressure is given in solid curve at bottom.

Atlantic and Pacific for example; they are notable in spring in eastern U. S., too.

The existence of "dynamic" lows and highs does not refute any principles of frontal or air mass *analysis*, however uncertain the theoretical explanations, but such phenomena must be recognized as additional processes to be given due consideration in forecasting.

It might be added that the semi-permanent sub-tropical anticyclones (e.g., Azores High) are also warm and dynamic in character; the moving anticyclones of middle latitudes often finally merge into them. However, the Aleutian and Icelandic semi-permanent Lows are more statistical than dynamic, due to frequent frontal cyclogeneses, though deep dynamic lows often stick there for days at a time.—*R. G. Stone.*

The rôle of the tropopause in the dynamics of extra-tropical disturbances

"J. Bjerknes has especially studied the interaction between waves in the Polar Front surface and waves of the tropopause, showing that the latter cannot be the primary ones. According to J. Bjerknes these induced tropopause waves mainly consist in a horizontal meridional oscillation of the air at the oblique tropopause.

E. Palmén adds to the above effect also the pumping effect of cyclones and anticyclones to which the stratosphere is subjected, using aerological data to show that the tropopause is often lower, the stratosphere temperature higher, in deep cyclones of our latitudes than in Arctic regions in winter. Such a state could not be attained merely by advection of the low and comparatively warm Arctic stratosphere, but the stratosphere must also be sucked down by the cyclonic vortex (and by analogy pushed up in anticyclones).

The author then proposes the following solution, which takes account of both effects and also satisfies the well-known statistical data from the upper air of Dines and Schedler. The *suction effect* of Palmén ought to predominate during and shortly after the most intense processes of cyclogenesis or anticyclogenesis ("Verwirbelung"=the transformation of a frontal wave into a vortex), when the suction effect of the "circular vortex" (according to V. Bjerknes, 1921, sinking in cyclones, lifting in anticyclones) will be most pronounced and a "new tropopause" has not yet had time to form at the normal height.—The *advective effect* of J. Bjerknes, on the other hand, will most likely predominate not only during the initial wave-like stage of Polar Front disturbances but also during their final stage, when the violent pumping effect of the "Verwirbelung" may have developed a kind of free oscillations of the tropopause, which liberate themselves from the tropospheric vortex and are propagated in the ordinary way eastwards.—These initial and final stages are of much longer duration and are also predominant in intensity as compared with the stage of "Verwirbelung" in those rather low latitudes from which the statistical data of Dines and Schedler were collected. This may explain why the statistics quoted speak in favour of the advective effect, whereas the intense high latitude cyclogeneses studied by Palmén show all the characteristics of the suction effect."—*T. Bergeron.*

VIII. THE TEPHIGRAM*

In order to forecast local showers and thunderstorms successfully it is necessary to interpret aerological data in the light of a reliable analysis of the synoptic chart. An individual sounding made in the lower troposphere is more immediately significant in summer than it is during the winter months. That is, in the warm months upper-air conditions immediately above a station are relatively important in determining the weather for the particular day, whereas in winter the rapid advection of air masses may so completely change the upper-air conditions over any given point that an attempt to forecast solely from the data in one sounding would be fruitless. Because of this control of the weather by local upper-air conditions, forecasters now consider upper-air soundings almost indispensable in summer work. In order to portray most effectively the state of the upper atmosphere, particularly from the standpoint of energy transformations, several diagrams have been suggested.

Hertz developed a diagram (later modified by Neuhoff) on which an aero-

*The procedures with the tephigram described herein may be readily applied to other thermodynamic diagrams such as the modified pseudo-adiabatic charts of the U. S. Weather Bureau (Upper Air Map C and forms 1147 and 1126 Aero.) described in Article III, p. 18, to the Refsdal emagram and "Aerogram", and to the Neuhoff diagram.

logical sounding might be plotted so that one could trace the path of any chosen particle of air as it ascended or descended through the surrounding medium, in this manner making it possible to see if work is being done by the rising air particle or if it is necessary to supply mechanical energy in order to make it rise. This diagram, which has long since found a niche in classical meteorology, forms the basis of modern attack on problems of the upper air through energy diagrams. The emagram and aerogram of Refsdal[2] and the tephigram of Shaw[3 4] have superseded the Neuhoff diagram because of greater simplicity in usage and adaptation.

Whatever the type of energy diagram used it is important to note at the start that in all of them it is assumed that one and only one element of air rises in some manner through the stationary environment. Thus the indications expressed in the diagram will depend in no small manner on the properties of the particular chosen particle. Actually the integrated effect of the large number of particles making up the air column should be considered. There is some doubt as to whether it is justifiable to neglect several other factors which may conceivably enter into the process. The ascent may be in the nature of lifting of an entire layer rather than the rising of a "bubble" of air, small compared with an air mass. Then there are the non-adiabatic effects of radiation. In spite of the complete neglect of such complicating factors, energy diagrams, based on the assumption that one relatively small mass of air rises through a more or less steady environment, have been used successfully. This fact *per se* indicates that the fundamental assumption is in the main justified. Furthermore, anyone who has had the opportunity to witness gliding activities can hardly doubt the existence of relatively small, rapidly rising up-currents, called by the gliding enthusiasts "thermals".

The tephigram, with which we shall here be concerned, acquires its name from the thermodynamic quantities that are its coördinates: temperature (T) and entropy (ϕ). The term entropy I shall not attempt to define here, for it is a concept which defies descriptive definition, and is a sort of mathematical adaptation. Its derivation may be found in any textbook on thermodynamics. Generally, the synoptic meteorologist regards entropy as something which is proportional to potential temperature, for it may easily be shown that:

$$\phi = C_p \log \theta + \text{constant} \qquad (1)$$

where ϕ is the specific entropy of dry air, C_p the specific heat of air at constant pressure, and θ the potential temperature. The choice of a value for the constant is entirely arbitrary, for, as with energy, it is not the absolute value of entropy that matters, but rather the changes.

Entropy is used as the ordinate of the tephigram. Because of the relation (1) and the ease of obtaining θ, the logarithm of the latter is often used as an ordinate in addition to entropy. The abscissa of the tephigram is temperature, usually expressed in degrees C. In the diagram shown in Fig. 13 the temperature increases from right to left; in some types of the diagram this scale is reversed (Fig. 15). The solid lines

[2]A. Refsdal: Der feuchtlabile Niederschlag, *Geofysiske Publikasjoner*, Vol. 5, No. 12, 1930; Das Aerogram, *Met. Zeit.*, Jan. 1935, pp. 1-5; *Geofys. Publ.*, vol. 11, No. 13.

[3]Sir Napier Shaw: Manual of Meteorology, Vol. III, The Physical Processes of Weather, Chapt. 7, Cambridge, 1930. Cf. Woolard et al: Graphical Thermodynamics of the Free Air, *Mo. Wea. Rev.*, Nov., 1926, pp. 454-457.

[4]C. M. Alvord and R. H. Smith: The Tephigram, M. I. T. Met. Course, *Prof. Notes* No. 1, 1929 (out of print). Repr. *Mo. Wea. Rev.*, Sept., 1929, pp. 361-369.

sloping downward from the left to right are lines of equal pressure. From Poisson's equation it is readily seen that these isobars are defined by the rectangular coördinates: potential temperature and temperature. In the diagram reproduced here (Fig. 13) the isobars are expressed in millimeters of mercury; more frequently the unit used is the millibar. Broken lines sloping slightly to the right of the vertical are lines of equal specific humidity (on some diagrams the mixing ratio is used instead), the values being the *saturated* mass of water vapor in grams per kilogram of moist air which can exist under the appropriate temperatures and pressures. The solid curved lines sloping upward from the left to right are pseudoadiabats — lines which represent the path of a rising saturated particle of air precipitating its moisture as soon as it is condensed. Note how the slope of these curves approaches the horizontal lines of equal potential temperature (the dry adiabats) at low temperatures, because of the decreasing saturation humidity content as absolute zero temperature is approached.

There are several ways to plot a tephigram, the most convenient depending upon the quantities one has at hand. For example, it is possible to use the rectangular coördinates, potential temperature and temperature; then again one may plot temperature against pressure by using for coördinates the slanting isobars and vertical isotherms. In Fig. 13 is plotted the sounding for Oklahoma City on June 20, 1935; this is the solid line, the significant points of which are small circles, A to H. In conjunction with the tephigram it is also helpful to construct what is known as a *depegram*—a curve showing the variation of dew-point, and hence of moisture, with elevation. This is constructed by plotting specific humidities at the significant levels against the corresponding pressures, which is more convenient than computing dew points and gives precisely the same result. The depegram for the Oklahoma ascent is represented in Fig. 13 by the dotted lines connecting crosses. From the tephigram one may easily determine the stability of any given stratum, for the horizontal lines, being lines of constant potential temperature, are dry adiabats; the vertical lines, isotherms; and the curves sloping upward from left to right are saturated adiabats. Thus the layer EF possesses a dry adiabatic lapse-rate, AD contains a temperature inversion (probably a ground radiation inversion), and the layer DE is in conditional equilibrium as the lapse-rate lies between the dry and the saturated adiabats. The depegram enables one to get a picture of the relative dryness of the various layers. For example, the surface layer at A is almost saturated, for here the dew-point is nearly equal to the temperature. Above the point E′ the air is very dry, a fact shown by the comparatively large distance between the temperatures at E, F, and above, and the dew-points at these levels shown at E′, F′, etc.

The most important use of the tephigram lies in indicating the amount of potential energy, in the overlying air column, which may be converted into the kinetic energy of a thunderstorm. For this purpose it is necessary to choose some individual particle of air and follow by means of the diagram the path it would take if subjected to vertical displacement up through the entire air column. It is assumed that the surrounding air remains at rest while this displacement of a unit element of air is taking place. It might be assumed, for example, that the point A is carried upward. In this event the particle of air represented by the point

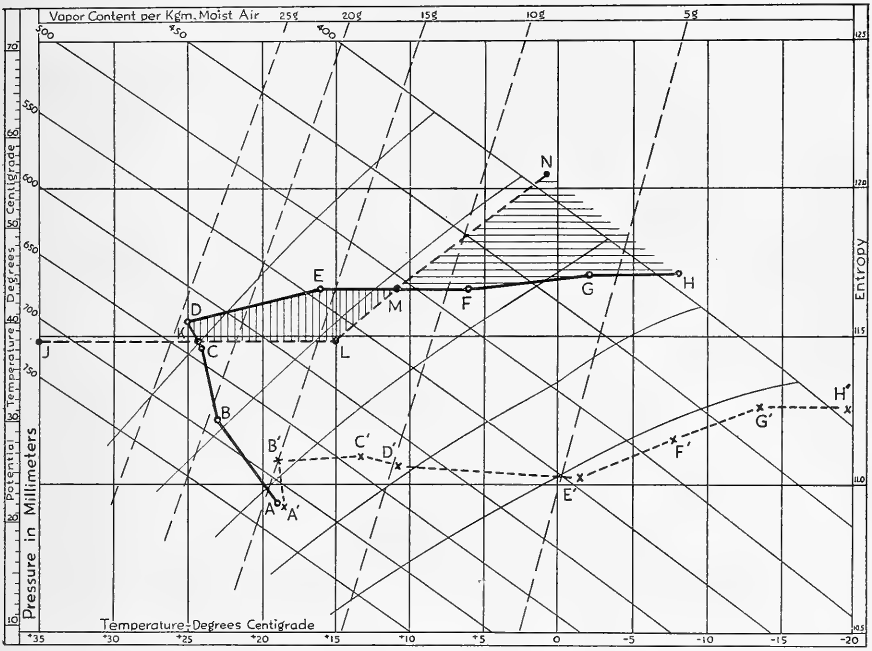


FIGURE 13. TEPHIGRAM FOR OKLAHOMA CITY, JUNE 20, 1935—EARLY A.M.

A would at first cool along the dry adiabat; that is, it would trace on the tephigram a horizontal line. But this could go on only until saturation occurred, for beyond this point the sample of rising air would no longer cool at the dry adiabatic rate but at the rate of expansional cooling for saturated air given by the pseudo-adiabats. With the help of the diagram this point of saturation is readily determined. As the particle ascends the mass of water vapor per unit mass of air (the specific humidity) remains constant until condensation begins. When saturation is reached the specific humidity of the rising particle is the maximum amount of moisture possible at that temperature and pressure. These lines of maximum (or saturation) specific humidity are given in the tephigram by the dashed lines which slope upward slightly to the right of the vertical. Consequently in order to find the point of saturation one has merely to trace a line horizontally from the originally chosen point on the tephigram until it intercepts the line of saturation specific humidity passing through the corresponding point of the depegram (that is, through the specific humidity of the original particle). Thus the horizontal line from A would continue to the right until it meets a line drawn through A′ parallel to the 15 g/kg line of saturation specific humidity. Beyond this point the rising particle follows the curved pseudoadiabats. This process will be clarified by selecting some other point from which a parcel of air would logically be more apt to rise through the air column.

The ascent plotted in Fig. 13 was made in the early morning; thus the large inversion in the surface layers is for the most part a radiation inversion. Under the effect of insolation as the day progresses the air near the ground may be expected to warm up considerably. Therefore the point A will be transferred to higher and higher temperatures while the pressure remains essentially the same. Thus A moves along the 720 mm isobar until the maximum temperature is reached. It is for the forecaster to decide what this value is most likely to be. In this case let us say that the temperature will rise to 35°C (95°F). A is thereby shifted along its isobar to J. Let us consider that there is no addition or subtraction of moisture so that the specific humidity remains constant at about 14 g/kg. Then as J rises its path is given by the dashed line; it follows the dry adiabat until its temperature falls to the value where the specific humidity (14 g/kg) is the saturation quantity. Thus J is moved horizontally until it intercepts the 14 g/kg moisture line at L. Beyond L the path of a rising particle is given by the pseudoadiabat LN.

The lapse-rate in the surface layer has been materially changed since the morning hours, and in view of convective mixing it is safe to assume that a dry adiabatic lapse-rate has been established up to the point K. Along JK there is neutral equilibrium with respect to dry air. That is, a particle of air along JK that is vertically displaced is neither assisted nor resisted by the density distribution. But while this adiabatic lapse-rate is built up, the structure of the air aloft, barring any change of air-mass properties, remains essentially unchanged. Thus above K, the point of intersection of the isentropic line and the original sounding, it is assumed that there are no major changes in air-mass properties. Consequently there still remains in the layer KD a portion of the original ground inversion. Now if a particle at J is forced to rise, following the path KL, its temperature at each level (defined by a particular isobar) would be lower than that of its surroundings. In fact, even after saturation at L the rising particle would remain colder than its environment until the point M when the temperatures of the rising air and its surroundings would become just equal. Throughout the layer from K to M work is required to lift a particle forced upward from K. The amount of energy required for this purpose may be shown to be equal to the vertically hatched area KLMEDK. On the original scale of the diagram from which Fig. 13 is reproduced, 1 square cm is equivalent to 2×10^5 ergs per gram; in Fig. 13, 1 square cm is equivalent to 3.3×10^4 ergs per gram.

Beyond M, a rising particle, following the pseudoadiabat MN at every stage in its ascent, would be warmer than the surrounding air. In this manner, after reaching M, it will rise of its own accord, so to speak, for it is now less dense than the air composing its environment. From energy considerations it may be shown that the energy liberated by a unit element of air rising from M to N is given by the horizontally hatched area MFGHNM.

(In practice it is customary to color in red the horizontally hatched area and in blue the vertically hatched area.)

To generalize:

Where the path of a rising particle lies above the tephigram of the sounding, energy is available for producing overturning, which may result in thundershowers, and the amount of this energy is given by the area (in *this case called positive area) enclosed between the path of the rising particle and the tephigram.*

When the path of a rising particle lies below the tephigram of the sounding, stability is indicated, and the area (called negative area) enclosed between the path of the rising particle and the tephigram represents the amount of energy which must be overcome if the particle is to penetrate the layer.

From these considerations it is clear that large positive areas and small negative areas are most favorable for the development of a thunderstorm. While there is no substitute for experience in working with the tephigram, it is possible to make some statements of a general nature which should assist the beginner in his use of the tephigram. Such a discussion, consisting of the indications and limitations of the tephigram, particularly when used in conjunction with a reliable weather map analysis, is taken up in the article on Thunderstorms, below.

A Note on Estimating Conditional and Convective Instability From the Wet-bulb Curve

Normand* has called attention to the fact that the usual method of indicating humidity by a depegram shows only the *variation* of humidity in a sounding, and that it cannot be used directly for visualizing the amount of energy available in a particle of air which rises from some particular layer. He pointed out further that in order to make reasonably correct deductions about the stability of such a particle of rising air, it is desirable to have a thermodynamic representation of humidity on the tephigram as well as one for temperature. In order to obtain such a representation he has advocated that the saturation (in practice, the wet-bulb) temperatures of a sounding be plotted on their appropriate isobars. This method is a natural sequence of the use of wet-bulb temperatures of the free air first advocated by him in 1921. He argued that the "saturation temperature" curve (S.-T. gram; or estegram) of a sounding should be used because the tephigram and depegram alone do not consider the actual state of humidity, and because it is desirable to have a curve which can be compared with the saturation adiabats just as the temperature-height curve can be compared with the dry adiabats. Drawing conclusions concerning the "liability" to instability from a tephigram alone by comparison of the latter with saturation adiabats is undesirable, for if the air is dry at all heights it is a waste of time to consider it as if it were moist. When that method is used it is the same as assuming that it is possible by a meteorological process to saturate dry air without altering its temperature! But it is well known that dry air which passes over a large lake or the ocean does not become saturated without a simultaneous change in its temperature. It is necessary when estimating the probable effect of an increase in humidity to assume a decrease in dry-bulb temperature, an increase in the dew-point, and little change in the wet-bulb temperature. This is the most important justification for including a wet-bulb or saturation temperature curve (these two curves are practically equivalent to one another) alongside the tephigram in estimating "liability" to instability.

Normand has found the addition of the wet-bulb temperature curve on the

* Normand, C. W. B.: Graphical indication of humidity in the upper air, *Nature*, 3rd October, 1931, Vol. 128, p. 583.

tephigram permits a more clear-cut classification of tephigrams according to vertical stability. In tropical regions the state of *conditional instability* generally prevails, and in middle latitudes it occurs rather frequently. Sometimes, however, it is associated with settled weather and sometimes with disturbed; in studies made in India the tephigram alone gives no criterion which allows a correlation with the type of weather. The Indian meteorologists found, however, that the important criterion in that connection is the vertical distribution of water vapor, and that a wet-bulb curve drawn beside a tephigram gives a more definite picture of the amount of energy that is likely to become available as a consequence of convection reaching the condensation level.

The use of estegrams together with tephigrams makes it possible to classify the conditions of particles of air with respect to their environment into three classes of *conditional instability*:–

1) Particles of air which if raised adiabatically will release more energy in the upper unstable portion of their ascent than must be supplied to them in their lower, stable portion, are in a state of *(real) latent instability*.†

2) A case in which a particle can be raised with a given supply of energy to a position where it is in an unstable environment and thus some of its energy can be liberated, but the energy so liberated is less than that supplied from below, is termed a state of *pseudo-(latent) instability*.

3) In the third case, when the lowest saturation adiabat tangential to the tephigram does not intersect the estegram, there is neither *latent instability nor pseudo-instability*. In other words, there is no energy realizable at all.

A particle of air has latent instability or pseudo-instability if the saturated adiabat through it (for example, the initial position of the particle under consideration) on the wet-bulb curve intersects the curve of the tephigram. The latent instability can be realized, however, only when the amount of work which must be done to raise a particle of air to the condensation level (from which it would thereafter rise freely) is less than the energy which will be realized during its further ascent (i.e., *only* in the case of *real latent instability*). If the saturated adiabat through the point of the initial wet-bulb temperature of the particle under consideration does not cut across the line of the dry-bulb temperature, then no upward displacement of the particle can bring it to a level where it will be as warm as its environment, so then the atmosphere is stable for all *particle* displacements—large or small.

In order to release the energy in air which is in a state of latent or pseudo-instability a trigger action is necessary, notably surface heating, and surface evaporation, but also lifting by a cold front, a mountain, or by convergence; release by lifting however, is strictly a case of convective instability in which of course wholesale release of conditional instability is an intermediate process (see below).

The recognition at a glance of the layers in which all particles have pseudo or latent instability is easy when the wet-bulb curve is entered beside the tephigram. For instance, in Figure 14 the layers with latent instability are determined by extending the moist adiabat which is tangential to the portion of the tephigram which approaches the wet-bulb curve most closely down to the surface

† Normand first defined the concept of latent instability in an article on tropical storms in *Gerlands Beit. z. Geophys.*, Vol. 34, p. 234, 1931; the principle is examined in detail by him in the *Q. Jn. Roy. Met. Soc.*, 1938, p. 47, ff.

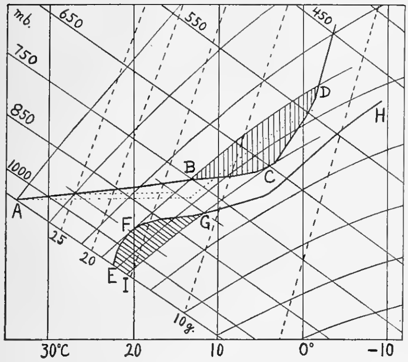


FIG. 14. A SCHEMATIC TEPHIGRAM SHOWING PRESENCE OF LATENT OR PSEUDO INSTABILITY for particles in the layer between E and G, which can be realized somewhere between B and D. The curve ACD is the temperature sounding; the curve EFGH is the wet-bulb temperature curve (estegram). The amounts of energy must be determined for any chosen particle in the usual manner; the shaded areas are not energy areas. EFGI merely marks the layer E-G, in which any chosen rising particle has real or pseudo-latent instability, which can be realized somewhere between B and D.

The paths of a rising particle from A and from F are dotted lightly. It is evident that the particle at F has much more latent instability than that at A, while a particle between F and G has only pseudo instability. Mixing and heating in the lowest layers would probably increase the dry and wet bulbs at A and E, so that in a forecasting problem one would draw the probable curves that would result from these effects and then consider the instability situation (see Fig 15).

level. Then all particles in the layer EG (i.e., where the wet-bulb curve curve lies to the left of CGI) have latent instability. Extending the moist adiabat which is tangential to the wet-bulb curve at F, indicates the layer in the environment aloft where any particles of air forced up from the layers between E and G would be able to realize their latent instability. Note, however, that the amount of latent instability (i.e., ratio of positive to negative energy areas) varies greatly for the different particles lying between E and G. One selects the particle which is likely to be heated, usually the surface one or a mean of particles near the surface, and draws the positive and negative areas for it in the manner described by Mr. Namias. However, the wet-bulb curve permits one to judge better how representative any given chosen particle will be for the trigger effects likely to occur (see figure 15).

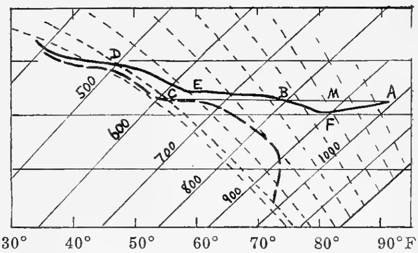


FIG. 15. LATENT INSTABILITY ON A TEPHIGRAM FOR A DAY WITH SQUALLS AND A DUSTSTORM AT AGRA, INDIA, MAY 22, 1929.—(A case from the paper by B. N. Sreenivasaiah: A Study of the Duststorms of Agra, *Memoirs of the India Met. Dept.*, Vol. XXVII, Part I, 1939.) On this date, a comparatively weak but rather prolonged duststorm occurred between 16:20 and 18:45 h, I. S. T., with well marked squalls at 16:20, 17:20, and 18:10 h. The synoptic weather charts showed that associated with the passage of a low pressure wave across north India, the seasonal low over northwest India became accentuated by the 21st morning and the pressure gradient over Sind and Baluchistan became marked. Numerous duststorms occurred in northwest India on the 21st; Agra also had one on that evening. The 22nd morning chart shows a marked trough of low pressure extending from Baluchistan to the west United Provinces hills; the low pressure wave apparently passed northeastwards subsequently. The duststorm of Agra on the 22nd was evidently associated with this spell of disturbed weather.

A meteorograph was sent up at Agra at 18:05 h (I. S. T.), i.e., actually during the period of the weak duststorm but after the first two principal squalls of the duststorm had occurred. There was a comparative lull in the phenomenon at the moment of the ascent, but within five minutes thereafter the last of the three squalls occurred. The tephi- and estegrams relating to the ascent are drawn. These show *real latent instability* for layers of air between the surface and about 460-mb level, the *environment of latent instability* (where it can be realized) lying above 690 mb. The tephigram shows the existence of a superadiabatic gradient of temperature between surface and 1-gkm level. The actual magnitude of this lapse rate is 13.5F°/gkm, whereas, according to the criterion of Brunt (Phys. Dyn. Met., p. 45), a lapse rate of 11.3°/gkm between surface and 1 gkm would be enough to cause (absolute) instability in the layer. Even up to 1.5 gkm conditions of lapse rate favorable for instability exist on this day, the actual rate being 10.7°/gkm, whereas the critical (neutral) rate is 10.3°/gkm. In any case, the layers up to 1 km are definitely unstable. These conditions would provide a "trigger" of sufficient magnitude to initiate displacement of the lower layers.

It is seen that a particle from A starting on account of instability in that layer will gain energy in the initial portion of the ascent up to B (794 mb). The gain is AFBMA. After B, the ascent up to D would involve a loss of energy BCDEB. It is only if the particle can reach D (500 mb), that it will come into the environment of latent instability and begin to gain energy; and it can reach D only if BCDEB is less than AFBMA. But actual measurement of these areas shows that BCDEB is slightly greater than AFBMA, being roughly in the ratio 14:11. This is what one would expect, if one remembers that the meteorograph was sent up only 5 minutes before the occurrence of the third squall, there being a comparative lull between the second and third squalls. During five minutes the meteorograph could have ascended only about 1 km. So, it is only up to about a km above ground that the tephigram can be considered to be representative of the conditions before the squall; above it, the tephigram shows conditions during the occurrence of the duststorm. As the temperatures rise in the upper levels after a duststorm, the area BCDEB is perhaps greater than it would be before the occurrence of the phenomenon; for the same reason the temperatures along BF may have been lower before the squall, making the area AFB slightly greater.

The possibility of descending cold air from a dust- or thunderstorm which occurred at neighboring stations having travelled to Agra and lifted up the "latently unstable" mass of air is not entirely ruled out, but in the absence of positive evidence to this effect and in view of the fact that the superadiabatic lapse rate existing is itself sufficient to cause the phenomenon, it seems reasonable to attribute the duststorm on this day to the trigger action of insolation leading to a superadiabatic lapse rate. The latent instability in this case is apparently built up by the moisture brought in by the passage of the low-pressure wave referred to above in the description of the synoptic situation of the day.

The fact that dust or thunderstorms do not occur more frequently than they do, although the vertical temperature gradients in the lower levels may be adiabatic or superadiabatic day after day in the summer months over Agra is probably due to the want of the necessary condition, viz., latent instability.

In this sounding all particles up to about 460 mb are latently unstable, an extreme case; the level of greatest latent instability is at about 880 mb, but since the convection is obviously taking place from the heated surface, the proper analysis is to consider the energy areas for a rising surface particle. The decrease of wet bulb below 900 mb indicates that a surface particle has considerably less latent instability than a particle say at 880 mb whose condensation level would be at 750 mb giving a smaller negative area and far larger positive area than for the surface particle. The same would be true for particles up to around 620 mb. Mixing in the lowest 300 mb layers will increase the surface wet-bulb, however, and tend to distribute the latent instability more equally among the different levels, in these layers. If this sounding were an early morning one, the latter consideration would certainly deserve some weight in estimating the realizable latent instability for later in the day.

In many cases, however, the wet bulbs are much lower, relative to the dry bulb, at the upper levels than in this sounding, with only a shallow moist layer (but not necessarily a temperature-inversion) near the surface which will often appear to have considerable real latent instability. But here it is well to allow for the effect of convection in lowering the surface wet bulb by mixing with the drier layers aloft, for in this way the amount of latent instability for the surface particles can be rapidly decreased before the convection reaches the condensation level. The method of assuming an average wet bulb for the lowest layers, in the same sense as Mr. Namias suggests using an average specific humidity, is often called for in such situations. On the other hand, evaporation often keeps the surface wet bulb from falling or even increases it in spite of mixing, and then the depth of the moist layer is increased at the expense of the drier upper air—which means the latent instability is increasing. Likewise advection of moist or dry currents aloft will alter the state, as explained in the article on Isentropic Analysis. Isentropic cross-sections and charts are not conveniently compared with estegrams since the former contain only θ, and q (or condensation pressures).

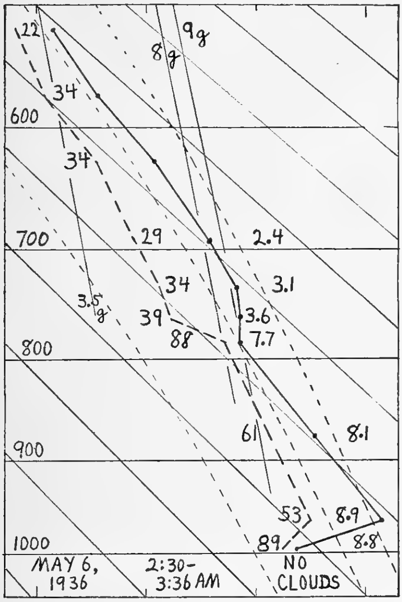


FIG. 16. AIRPLANE SOUNDING AT MURFREESBORO, Tenn., plotted on the pseudo-adiabatic chart. The temperature curve is drawn as a full line with specific humidities written in to its right. The wet-bulb curve is drawn as a heavy dashed line with relative humidities entered beside it. This case is of interest because it shows conditional instability in most layers but without real-latent or pseudo-latent instability. There is a little convective instability aloft (e.g., just above 785 and 630 mb) where the wet bulb lapse rate is greater than the moist adiabat.

Bodily lifting of the surface layers by 100 mb or more would make some fairly thick Cu clouds whereas considerable solar heating of the surface layers will produce only Fc or Sc, if any clouds at all, (because the condensation level will be raised).

The sounding is in general rather stable in spite of moderate to high relative humidities and conditional instability in the lower levels; this is because there is no latent instability in the first km or two.—*R. G. S.*

In a study of some Indian tephigrams, Sohoni and Paranjpe** have applied Normand's suggestions and correlated their findings concerning the presence or absence of latent instability in the soundings with air-mass type and weather. The estegrams corresponding to the tephigrams were drawn in each case as follows:—the dry adiabatic (isentropic) is drawn through the point at the surface level, and then the isohygric (constant mixing-ratio line) through the corresponding dew-point is drawn. The saturation adiabatic which runs through their point of intersection meets the isobar at the surface level at a point which has a temperature coordinate of T° A. This is the adiabatic wet-bulb temperature of the air particle at the surface level. The wet-bulb temperatures at the other levels are obtained in the same manner, and then the estegram is drawn by joining all of the points so obtained.

Sohoni and Paranjpe found that the absence of latent instability is associated with dry, fine weather with occasional high clouds of non-convectional type, and latent instability with convectional types of clouds. Later studies in India confirm this and the wet-bulb curve is now widely used.

It is also practicable to determine the presence of what Rossby calls "convective (or potential) instability" of a layer of air by use of the wet-bulb temperatures on a tephigram (also on the pseudo-adiabatic chart). In any layer of air in which the lapse rate of the wet-bulb temperature exceeds the moist adiabatic rate, the wet-bulb potential temperature, and, therefore, the equivalent-potential temperature decreases upwards. A layer of this type is convectively unstable according to Rossby's definition. If such a layer is raised bodily and adiabatically in the atmosphere *without mixing with its environment or distortion* and in such a way that the difference of pressure between the top and bottom of the layer remains constant, then *it can easily be seen that every layer in which the wet-bulb lapse rate is steeper than the saturated adiabatic must ulti-*

** Sohoni, V. V. and Paranjpe, M. M.: Latent instability in the atmosphere revealed by some Indian tephigrams, *Mem. India Met. Dept.*, Vol. XXVI, Part VII, 1937, pp. 131-149.

mately become unstable if raised to the condensation level. Since energy is required to raise the layer, however, and the amount which must be supplied for this purpose may exceed the energy which can be expected from lifting or heating, the energy of convective instability may be unrealizable. The only important types of convective instability are those in which the energy which has to be supplied in order to lift the layer to the condensation level is less than the amount which will be realized from the resulting instability. *Such a gain in energy is most likely to be realized when the layer of air under consideration is originally nearly saturated or when latent instability is present from the start.* There may be convective instability with or without conditional instability *at the start*, of course. The use of the wet-bulb curve serves to classify the convective instability according to whether there is latent-, pseudo-, or no conditional-instability present at the same time, and hence the relative probability of its release by any trigger actions that can be foreseen on the weather map. In tropical regions and in summer in higher latitudes this simple method of estimating all types of instability on one diagram (tephigram, Neuhoff diagram, or emagram) will certainly appeal to practical meteorologists. Even on the Stüve pseudo-adiabatic chart the wet-bulb curve will be qualitatively helpful, in case there is insufficient time to plot an energy diagram.

Bleeker has shown that the adiabatic wet-bulb temperature is not strictly a conservative element for an evaporation process while the isobaric wet-bulb is not conservative for wet or dry adiabatic processes, but for estimating instability this makes no practical difference. Petterssen has lately suggested drawing the wet-bulb temperature curve on the pseudo-adiabatic chart to indicate convective (or potential) instability (see Fig. 16). This is a more convenient procedure than plotting a Rossby diagram, but the deviation of the wet-bulb curves from the moist adiabats is usually so small that the layers with convective instability cannot be read off so accurately as on the Rossby diagram, though this is not serious except in borderline cases when the inaccuracy of aerological measurements would probably not permit a definite conclusion anyway. —*R. G. S.*

IX. SYNOPTIC ASPECTS OF THE THUNDERSTORM†

In the preceding article the plotting of the tephigram and a partial interpretation of an individual sounding was outlined. This article will present briefly the probable causes of thunderstorms, indicating the use of aerological data in their analysis and forecasting.† However, the origin and development of the thunderstorm from the synoptic viewpoint are by no means fully understood as yet, and therefore offer a fertile field for research.

Perhaps the most convenient classification of thunderstorms is based upon the physical factors which presumably cause them:

1. Air-mass thunderstorms:
   - (*a*) From local convection.
   - (*b*) In thermodynamically cold air masses.
2. Frontal thunderstorms:
   - (*a*) Associated with a cold front.
   - (*b*) Associated with a warm front.

†Further discussion of thunderstorm forecasting will be found in Article X on Isentropic Analysis.

(*c*) Associated with an occluded front.
3. Orographic thunderstorms.
4. Thunderstorms in horizontally converging air currents.

I. Air Mass Thunderstorms

A. *From local convection.* The type of thunderstorm known for a long time as "local," "heat," or "local convective" may be singled out from other types in that it represents a form of penetrative convection due primarily to insolational heating in the lower layers of the atmosphere. No apparent frontal activity is involved, and the original cause appears to be purely thermal in nature. In summer these thunderstorms may be observed in almost every type of air mass. They are, however, by far most frequent in TG air, because the TG air has the most favorable moisture and temperature distribution with elevation. Steep lapse-rate and high specific humidity are most favorable for the development of Cu clouds into tall Cb. The tephigram under such conditions generally exhibits a large positive area and a comparatively small negative area. As pointed out in the preceding article, it is important to use critical judgment in selecting which particle of air will be assumed to penetrate the upper strata; furthermore, it is necessary to assume the temperature and moisture that particle will probably have before it ascends. Thus if one should choose the surface air particle having the temperature and moisture observed in the early morning hours he would, owing to the stability of the ground inversion, always obtain tremendous negative areas. In forecasting a maximum temperature for the surface particle one should be familiar with the locality for which he is forecasting. It is common practice to take for the maximum temperature of the day the surface temperature which would be potentially equivalent to that at the top of the ground inversion. The physical reasoning behind this procedure is that as insolation warms the ground the stability of the overlying inversion precludes any appreciable upward transfer of the heat supplied the surface layer. Consequently, the lowest layer warms up rapidly until the ground inversion with its stability is completely wiped out. After this takes place most of the excess heat is carried upward by convection, and the temperature at the surface remains at a fairly constant maximum. In choosing the specific humidity of the particular element which is to rise through the surrounding air mass it is customary to assume that the specific humidity in the surface layers remains constant. This is logical because, exclusive of evaporation from the surface and mixing with the air above, the mass of water vapor per unit mass of air should remain sensibly constant in one and the same air mass; in a case where the specific humidity decreases rapidly upward from the surface, the mixing by convection and mechanical turbulence will lower the specific humidity of the surface air. In this case it would be erroneous to presume a constancy of moisture content during the day. In practice I have found it helpful in these cases to choose for the probable specific humidity of the rising particle the *mean* value of the observed specific humidities through the lowest layers up to the condensation level, or to the top of any ground inversion or stable layer having relatively high humidity.

A not uncommon case presents itself when above the moist surface air, presumably TG, there flows a current of warm dry air (Ts); if the TG air is shallow, the construction of a tephigram with the assumption that the surface particle will penetrate the Ts air will often be misleading, for it suggests there are large amounts of available energy (*positive* area) which actually, however, cannot be realized because of the mixing of this moist stratum with the dry air above. But generally, especially when the TG air is about 1 km deep, the tephigram gives a fairly true picture of the available energy, indicating (after extrapolation of the surface point) positive areas at the surface and aloft separated by a negative area. This stable layer (the negative area) tends to resist convection and must in some manner be penetrated by the rising particle before the Cu clouds can develop into thunderstorm proportions. It is conceivable that in some cases the extra energy needed to overcome stability may be supplied by the kinetic energy of the rising air —this energy causes the cloud to rise beyond the level at which it has the same density as its surroundings. The energy may at times be supplied by the lifting action of an incoming front or by orographic upthrusting.

In a special study of aerological material Willett† found that the TG layer, whether or not a front enters the synoptic field, must be at least some 3½ km thick for thunderstorms to develop. As he points out, however, his data contained no ascents made during June through September—just the time of the year when convective activity is the greatest. It is probable, then, that this conclusion does not hold for summer situations.* It is also likely that soundings made in or near a Cb cloud are not representative of the properties of the air mass in which the convective overturning is taking place, because these turbulent ascending bodies may be virtual "fountains" of moist air. Further research is required here.

In making use of the tephigram in forecasting local thundershowers it is important to consider the chronological *changes* in the positive and negative areas. Rapidly increasing positive areas and decreasing negative areas are highly indicative of future thunderstorm activity. If a succession of six- to twelve-hourly ascents is not available, it is very helpful to choose for comparison tephigrams of ascents made the preceding few days at other stations in the same air mass, preferably ones lying in the general trajectory of the air mass—that is, in the probable path of the air current before it reached the station for which the meteorologist is forecasting.

The local indications of convective thunderstorms are so commonly known[2] that we need not dwell upon them here in detail; some may be mentioned: the presence and growth of towering Cu during the day, the warm, muggy and often stagnant atmosphere, and the steep pressure fall by afternoon. The distribution of the Cu is frequently indicative, for it has been shown by Bjerknes[3] that tall Cu separated by fairly large clear spaces (*Cu. castellatus*) are most

†Discussion and Illustration of Problems Suggested by Analysis of Atmospheric Cross-Sections, M. I. T. and W. H. Ocean. Inst., *Papers in Phys., Ocean. and Met.*, Vol. 4, No. 2, 1935.

*For further discussion of this see my article in the Jan. 1938, *Bulletin Am. Met. Soc.*, p. 11.

[2]Cf. C. F. Brooks: The local, or heat, thunderstorm, *Mo. Weather Rev.*, June, 1922, Vol. 50, pp. 281-284. (Includes "Questionnaire [of 16 questions] for predicting local thunderstorms.")

[3]Physikalische Hydrodynamik. Berlin, 1933; Hydrodynamique Physique. Paris, 1934 (3 Vols.). In *Q. Jn. Roy. Met. Soc.*, April 1938, pp. 325 ff, Bjerknes shows that Cu convection does not release sufficient kinetic energy (motion) to cause Cu to develope into Cb unless the spacing—thickness ratio of the Cu towers is below a certain critical value.

favorable while flat Cu (*Cu. humilis*) covering most of the sky are contra-indicative of thunderstorm activity. There is, however, a limit to the narrowness of Cu, for tall narrow clouds are apt to be torn apart and dissipated by turbulence and wind shear. Thus there is an optimum size and space distribution for Cu in order for Cb to develop.

Although the following rule is more or less obvious, it is important enough to italicize: *Before attempting to use the indications of a tephigram in forecasting local convective showers make certain that no fronts will enter the field within the period for which you are forecasting.* Thus the intelligent use of upper air data goes hand in hand with a reliable analysis of the synoptic chart.

B. *Thunderstorms in thermodynamically cold air masses.* In this type of thundershower the steep lapse-rate in the surface layers is established chiefly by the transport of a cold (usually Pc air) current over a relatively warm land surface. The Pc air mass is considered cold in a thermodynamic sense, because it remains continually colder than the surface over which it is travelling. Consequently, the lapse-rate in its lowest layer must be steep. The situation is characterized in winter by instability snow flurries, in spring and early summer by instability showers which infrequently develop into thundershowers, and then usually mild, for the necessary ingredients of a well developed thunderstorm, air with high moisture and heat content, are lacking.

In summer time this kind of thunderstorm is not common, its formation apparently being hindered by the gradual transformation of the Pc properties as the air mass moves southward over warmer surfaces. In the spring, on the other hand, the Pc air may be transported from its snow-covered source region to the warmer bare-land surface in such a short time that the upper layers have little time to warm, and consequently very steep lapse-rates are established.

Another restraining influence on the formation of this type of thunderstorm is the presence of surfaces of subsidence within the cold air, which act as "lids" that stop the upward growth of the Cu clouds in the cold air; it requires considerable energy for the rising air to penetrate these stable surfaces. The limiting subsidence layers generally lie within the cold air in the shape of a vast, flat dome—the top being near the center of maximum pressure rise (anallobar), the edges becoming lower as the front boundary of the cold air mass is approached. Consequently, for some distance behind a cold front conditions are unfavorable for the type of shower in question, for here the surface of subsidence is lowest. Moreover, near the front the stability of the frontal (vertical) transition zone helps to check the convection in the cold air. (See Figs. 17 and 18.)

It is difficult to apply the tephigram to these situations. One cannot make certain of the extent of the warming or humidification of the lowest layers of the air mass, or for that matter, the changes in structure which may go on aloft. A sequence of tephigrams constructed from flights at fairly short intervals is the most helpful. Lacking these, perhaps the best indications are expressed in the history of the cold air mass traced through analysis of the surface charts.

II. Frontal Thunderstorms

A. *Associated with a cold front.* Cold-front thundershowers are the most interesting and generally the most violent of all. In the older terminology they are the "line squall" thunderstorms. The predominating cause appears to be the mechanical upthrusting of the warm air by the advancing cold wedge. The indications for thundershowers with the passage of a cold front are to be found largely in the upper air characteristics of the warm air mass and in the structure and movement of the cold front. The tephigram is well-adapted to these cases. The soundings used should naturally be those in the warm sector. As before, a sequence of soundings at one station is best, but if this is not possible it is advisable to study several tephigrams that show the trend of the change in the characteristics of the warm air mass. Here again, increasing positive areas and decreasing negative areas are most favorable indications for thunderstorms, particularly if the current tephigram shows appreciable positive areas. The most indicative lapse-rates are conditionally unstable. Convective instability is also generally indicated, for it may be that the lifting of entire layers is responsible for the genesis of the showers.

In forecasting cold-front thundershowers it is important to consider the diurnal change in lapse-rate. In the early morning hours there is usually a ground inversion present, while in the afternoon the surface layers are characterized by a superadiabatic lapse-rate (Fig. 19). Then the air mass is in its most favorable state for producing thunderstorms once the "trigger action" of a cold front is supplied. It is for this reason that cold fronts passing during the night are frequently accompanied by no thunderstorms, while the same fronts, passing during the afternoon or early evening, are characterized by thunderstorms all along their length.[4]

The cold air immediately behind the front is cooled by evaporation of the falling rain; in this manner the sharpness of the front is maintained.

The air masses preceding the front on which cold-front thunderstorms may be observed hardly seem to be restricted to any individual types, although they occur with a pronounced maximum frequency in TG currents, and only rarely (if ever) in the dry currents which move from the southwestern U. S. and Mexico.

Cold-front thunderstorms may be further classified as prefrontal, frontal and postfrontal, according as they occur appreciably before, at, or after the front passage. We need spend little time here on the fine points of this classification. Prefrontal types may be due to mechanical uplifting of the warm air some distance ahead of the cold front, or to the entrapping of the warm, moist air below the overrunning squall head. In the latter case it is conceivable that extremely steep lapse-rates and violent storms may be brought about. Those thunderstorms occurring immediately at the front are presumably the normal type due to uplifting. The postfrontal thunderstorm probably belongs, for the most part, to the class discussed as "thunderstorms in thermodynamically cold air masses."

An interesting situation occurs when a cold front slows up in its movement and becomes quasi-stationary. Then thunderstorms may be occurring at various points along the front and persisting for a relatively long time. These are associated with small wave disturbances moving

[4]This is brought out clearly in the statistical frontal study of Thunderstorms in Ohio during 1917, by W. H. Alexander, C. F. Brooks and G. H. Burnham, *Mo. Weather Rev.*, July, 1924, Vol. 52, pp. 343-348.

along the front and supplying the necessary "trigger action" to release the energy in the warm air. The showers apparently form both in advance of and behind the front, but the structure and action of these waves is not fully understood. Excellent examples of these quasi-stationary frontal thunderstorms can be observed in summer over New England.

B. *Thunderstorms associated with a warm front.* Thundershowers associated with a warm front are not nearly so common as those with a cold front, the reason being that the vertical motions brought about by a warm-front surface are not as pronounced as with a cold front. Nevertheless, in favorable circumstances one may observe on the surface weather map the outbreak of a large number of thunderstorms definitely associated with a warm front. It appears that this type is almost entirely confined to fronts where TG is concerned, particularly in autumn, winter and spring; there are occasional summer cases, where the warmer air mass is a transitional polar current whose temperature and moisture content have been increased by stagnation over the southeastern U. S. almost to the characteristic values for tropical air. These warm-sector air masses capable of producing warm-front thundershowers are well supplied with potential energy, which is on the verge of being released. The releasing agent or "trigger action" necessary for the transformation of this potential energy into the kinetic energy of a thunderstorm is the upward deflection of the warm air over the cold wedge.†

Another important factor is the horizontal convergence of air in the warm sector; this serves to steepen the lapse-rate in the warm air, for the surfaces of equal potential temperature are stretched vertically.* This process may be considered the reverse of subsidence, wherein horizontal divergence occurs and the lapse-rate becomes more stable. With horizontal convergence in the warm sector, the air involved is so transformed that the relatively small vertical motions supplied by ascent over the cold wedge are sufficient to develop thunderstorms. It is obvious that this same conditioning process is also important in setting the stage for other types of thundershowers.

The major action in warm-front thundershowers takes place aloft. The storms are usually less violent than other types. Furthermore, the characteristic cloud display may be obscured by the warm-front cloud deck.

The diurnal variation in frequency of these showers is not as pronounced as with the other types, particularly in the non-summer months, when they seem to be about as frequent at night as by day.

Just as with quasi-stationary cold fronts, waves may form on a slowly moving warm front and may at times be responsible for the development of thunderstorms along it. It is not rare to find a comparatively sudden outbreak of violent showers along an east-west front separating TG from PC air, and usually this can be traced to some wave action which has released large amounts of indicated (on the tephigram) energy in the TG air.

In using the indications of the tephigram for forecasting warm-front showers one must locate accu-

†In India it has been shown that the cold air descending out from under one thunderstorm often spreads out along the surface as a local cold wedge and acts as a trigger to set off new thunderstorms nearby.—*Ed.*

*This is true only in case the original lapse rate was stable, otherwise convergence would tend to stabilize the air.—*Ed.*

rately the position of the warm front at the surface, and obtain a fairly good idea of its structure. A sounding made some distance in advance of a warm front, if subjected to the ordinary analysis for local convective showers, will rarely show thunderstorm indications, because here the erroneous assumption is made that the surface cold air, in reality belonging to the cold wedge, will rise through the upper strata. The proper procedure in these cases is first to identify the discontinuity surface separating the two air masses aloft, then instead of a surface particle assumed to rise take some element in the warm sector, ascribing to it a probable maximum temperature as in the case of air-mass convective showers, and then plot the ascent of this element on the original tephigram. It is best to choose for the rising particle the air at the base of the warm current.

C. *Thunderstorms associated with an occluded front.* Perhaps the largest number of summer thunderstorms are associated with fronts previously occluded, which, through their action on the field of flow, have regenerated into active cold fronts. It is important in summer for the analyst to follow closely *all* occluded fronts, for it is the rule rather than the exception for bent-back occlusions, weak as they may appear, to develop into surprisingly active cold fronts. It is largely the activity of these occlusions that makes summer-map analysis so interesting. They form, as do ordinary cold fronts, a pronounced field of convergence, making the front sharper and conditioning the preceding warm air to the point where it becomes rich in convective energy. In general, for purposes of thunderstorm forecasting, these cold-front occlusions may be treated in the same manner as cold fronts.

The occluded warm front is interesting in that it represents an upper cold front marching into the field aloft, well in advance of the surface warm front; the warm air originally above the warm front surface is thus displaced by a current of colder air. In this way, the lapse-rate above the lowest part of the warm front is made steeper. These warm-front occlusions seem to be characteristic over the southwestern plains, where the TG air is colder than other warm air masses (often NTP) which therefore override it, while NPP air, warmer at the surface than the TG, but colder than the NTP, rapidly overruns the TG and displaces the NTP; the warmest air is then pocketed between the TG wedge and the NPP wedge.[5] Thunderstorms are at times associated with this structure, though their exact mode of formation is not as yet definitely known; but it appears that the steep lapse-rates observed aloft in the NPP air mass play an important part once some particular body of air starts to rise. It is possible that insolational heating in the cool wedge (TG) may give the initial impulse, while the stability of the frontal boundary may be insufficient to prevent its penetration into the more unstable air aloft.

In making use of the tephigram in these situations one must consider the possibility of rapid change in structure of the upper air and must be familiar with the characteristics of all the interacting air masses.

III. Orographic Thunderstorms

Practically all types of thunderstorms are more frequent over hilly and particularly over mountainous

[5]Cf. H. Wexler: Analysis of a Warm-front-type Occlusion, *Mo. Wea. Rev.*, July, 1935, pp. 213-221.

country than over flat terrain. Besides there are many situations in which thunderstorms form only in that part of an air mass overlying certain mountain regions. The reason for this seems to be that rough terrain increases vertical turbulence, and slopes force marked ascent of the air. These upward deflections may be enough to set free any large stores of energy available in the air masses (especially the conditionally unstable). Solar radiation on mountains is more intense, and it is also conceivable that the higher angles of incidence of the solar rays on some mountain slopes may favor convective thundershowers.

A characteristic of many of these showers in mountainous regions is their tendency to remain almost stationary, presumably because the upward deflection of the air flow over the mountain is operative in only certain localities (windward sides) while dissipating forces are at work in others (leeward).

Another type of orographic thunderstorm is the coastal shower in conditionally unstable maritime air masses, frequent on the Pacific northwest coast of the U. S., where the ascent of fresh Pp air up the coastal slope and ranges supplies the "trigger action." A similar phenomenon occurs in conditionally unstable (or convectively unstable) Tg air masses invading the country along the Gulf coast, where the slight elevation of the land and initial heating over land appears to be sufficient to produce showers.

IV. Thunderstorms in Horizontally Converging Air Currents

It has been pointed out that horizontal convergence of air, through its vertical spreading action, makes the lapse-rate steeper. This process is undoubtedly present in most open warm sectors. There are many cases in which a bent-back occlusion in some manner establishes behind it a strong convergent field, so that the front appears to be elongating itself. In the U. S. this convergent zone is generally accompanied not only by an abrupt wind-shift, but also by typical cold-front characteristics; and there presumably are cases of such zones of convergence in which there is a pronounced wind-shift but no discontinuity in temperature. Such a zone cannot be called a front. The Norwegian meteorologists have found several cases of this kind over Europe.[6] The characteristics of such a trough of low pressure are similar to a front. Owing to the convergence there is frequently precipitation. As far as I am aware, no one has published an analysis of such a case in the U. S. It seems probable that bent-back occlusions here almost invariably regenerate into cold fronts, because the air masses concerned (usually Pc or Npp) seem to undergo modification along their trajectory more rapidly than the air masses over Europe (moist maritime-polar currents) (see Petterssen: Contribution to the theory of Frontogenesis).

Another type of convergence, which might have been discussed under the heading of occlusions, is that forcibly brought about by the rapid occlusion of a cyclone at its center. This occlusion may eventually lead to thunderstorms but only when the cold front rapidly cuts off the warm sector and the warm air is not very stable. This kind of thundershower is often observed over southern New Eng-

[6]J. Bjerknes: Investigations of Selected European Cyclones by Means of Serial Ascents (Case 3: Dec. 30-31, 1930), *Geofysiske Publikasjoner,* Vol. XI, No. 4, 1935 (2:00 Kr.).

land, the winter thunderstorms of that section being almost entirely of this type; rapidly intensifying cyclones (often secondaries) bring a tongue of Tm air to their centers where it is rapidly occluded, causing thundershowers along the coast. In the warm months of the year the effect of the colder water tends to retard the warm front, while the cold front proceeds to occlude the warm sector.

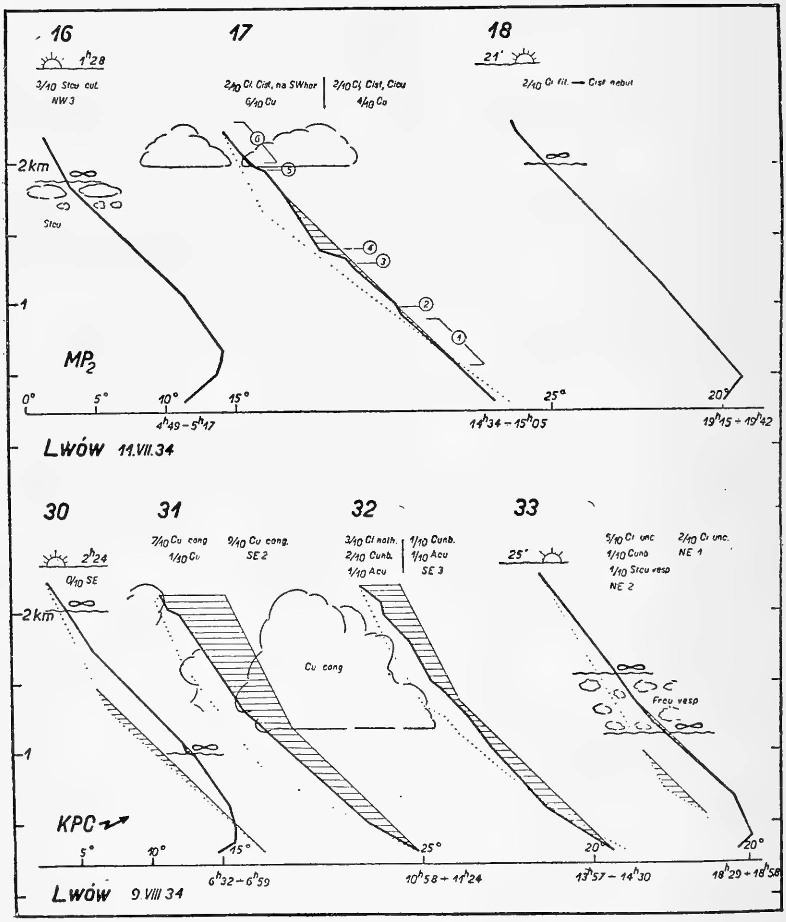


Fig. 17. Airplane Soundings at Different Hours of the Day, showing development of Cu clouds from insolation in a dry and a moist air mass. These were made at Lemberg (Lwów) Poland on July 11, 1934 (dry) and August 9, 1934 (moist), and are from the remarkable series of illustrations of convection published by Kochanski (*Comm. Geophys. Inst. Univ. Lwów*, vol. 8, No. 109, 1936). The soundings are plotted on Refsdal's emagram coordinates (similar to Stüve's diagram, i.e., log pressure vs T) and the areas of positive energy are shaded in the same manner as on a tephigram. The dotted lines are the temperature curves for the descent of the plane, which

show usually considerable difference from the ascent (times of beginning of ascent and end of descent are given at bottom). The clouds are realistically pictured, and their amounts and character listed at the top, also direction and velocity. On July 11 PM air from the NW has been moving over the station for almost 2 days. The 4:49 a.m. sounding made 1 hr and 28 min after sunrise, shows a nocturnal radiation inversion, and a few shallow Su clouds under a stable layer and haze line at about 1850 meters. At 14:34 h the clouds had thickened to .6 Cu with base at 2 km, the lapse rate averaged dry adiabatic but irregularities in some layers are evident, probably due to descending and rising currents, thermals). At 19:15 h (just 20 min before sunset) the clouds are gone, only the haze line remains, and the nocturnal inversion is already beginning, with a slight stability developed at all levels. That was a relatively dry air mass, like NPP in central United States.

The case of Aug. 9 was in cPW (i.e., returning NPC air of a Russian High) and there was enough moisture for instability to produce showers in the afternoon. At 6:32 a.m. (2 hrs 24 min after sunrise) there are only haze lines at 2 km and 1 km, at the bases of the slightly more stable layers. The descent, however, shows a much steeper lapse rate with the inversion wiped out and a positive energy area already in evidence. By 10:58 h the sky is full of Cu *congestus* starting at 1200 meters and large amounts of positive energy are present from the surface up to high levels. The descent was even more unstable. At 13.57 h Cb have developed with large cirrus anvils observed, the cloud base is higher, and the positive energy areas reduced by the convection working up to higher levels. At 18:29 h, 25 min before sunset, the clouds have largely evaporated and the remnants spread out into a shallow Fc *vesperalis* layer. Stability is returning rapidly along with a general subsidence (compare descent with ascent) and a strong radiation inversion has already begun to form. At higher levels and in the distance scattered remnants of Ci and Cb systems are still visible.

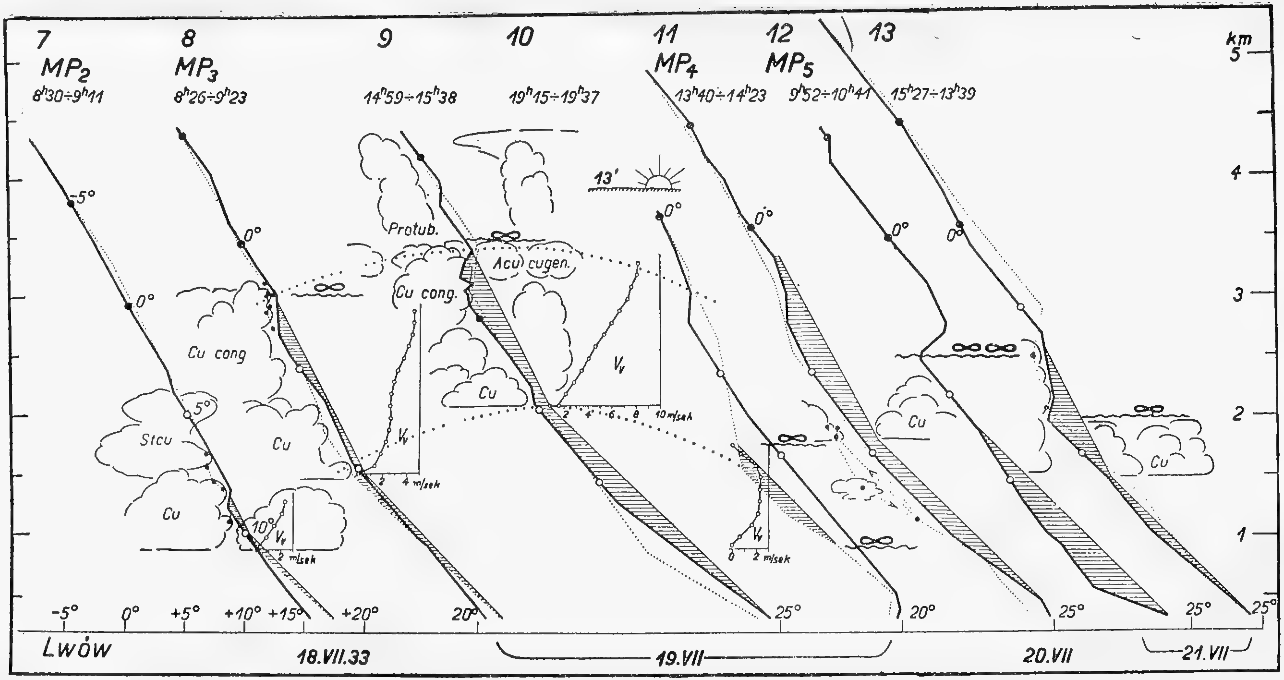


Fig. 18. The above diagram of a series of consecutive soundings is also taken from Kochanski's valuable work (ref. in legend to fig. 17) and gives a beautiful illustration of the sequence of events during passage of a summer high pressure area described by Mr. Namias on page 63, col. 2. (The legends are the same as on Fig. 17.) Maritime polar air invades southern Poland for over 6 days (July 17-21 +, 1933). In the sounding on the 2nd day it has been over Lwów, at 8:30 h, the air is rather stable and cool with fair weather Cu clouds at 1-2 km, the tops flattening out into Stcu. (The little inset diagrams by this and other soundings, marked "V_v", are the vertical velocities in m/sec in the clouds, calculated from the instability energy.) On the 19th three soundings were made, showing the diurnal course of events aloft in striking fashion. At 8:26 a.m. the Cu cloud base is higher, the air

warmer and the clouds thicker (now *congestus*) than on the 18th. Positive energy reaches to 3 km but the upper levels are rather stable yet. A small stable layer and haze line marks the top of the cloud level. (*Note*: white circles on the temp. curves are points above freezing, black dots are those below 0°C.) At 14:59 h the lapse rate is superadiabatic to over 1 km, the cloud base much higher than at 8 h, the warm cloud layer somewhat less thick because the upper stable layer has not been pushed up as much as the condensation level, though towering Cunb with anvil protuberances here and there poke through the stable layer (by penetrative convection) to nearly 5 km. Showers occurred in the vicinity.

By 19:15 h, just a few minutes before sunset, clouds have evaporated and the layers in which the clouds had been have settled down to their morning levels (as shown by dotted curves connecting the 3 soundings). On the 20th and 21st the increasing subsidence of the upper levels of the air mass chokes off the cloud tops under marked subsidence inversions (haze lines), in spite of considerable positive energy in the lower layers.

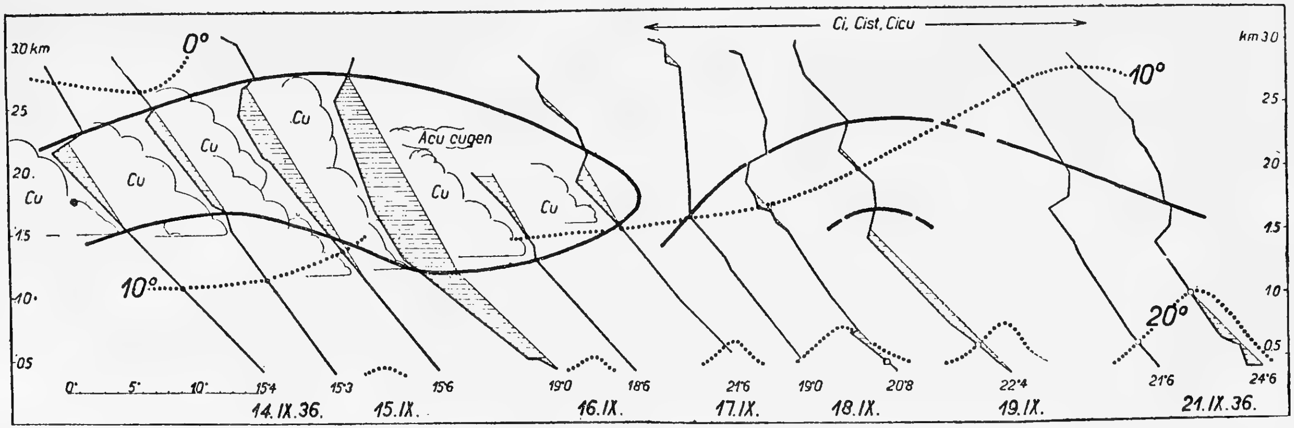


FIG. 19. Airplane soundings made once or twice daily (8-10 h and 10-16 h) during passage of a high pressure area (14-21 Sept., 1936) over a 1200-m deep mountain valley in the Beskid mountains of Polish Silesia. As analyzed here on emagrams by Kochanski (*Comm. Geophys. Inst. Univ. Lwów*, vol. 9, no. 115, 1937) attention is directed to the instability up to 2.7 km on the first 3 days, with Cu clouds capped by inversions (area enclosed by heavy line); and on the remaining days to the pronounced stability aloft, the subsidence inversions (heavy line) descending to lower levels and along with the warming up from below preventing any clouds forming. The domed structure of the subsidence inversion in the high is indicated (17-21st). Note the diurnal and day to day changes in elevation of the isotherms of 0°, 10° and 20° C shown in dotted lines, which indicate the manner of warming up of the air mass.

The Ice-Nuclei Theory of Rainfall

When a cloud top grows a considerable distance into region above the 0°C level, more and more ice crystals will form, creating a layer of water and ice particles and perhaps still higher one with ice crystals only. Bergeron developed a theory (later elaborated by Findeisen) that *large-dropped rain* will not be precipitated in considerable amounts until the top of the cloud is partly composed of ice crystals, because the difference in vapor pressure over ice and water leads to rapid growth of the ice crystals by evaporation from the drops, the ice crystals then falling out, melting and taking on more water on the way down as rain. Other clouds are more colloidally stable and give only fog or drizzle. This theory seems to agree with observations in middle and high latitudes, but is not yet accepted as universally true, as there is much evidence from the tropics and subtropics that certainly it is not always applicable there.

If and where the principle holds, however, then the most favorable conditions for the development of thunderstorms would exist when the distance between the lowest level where convective energy would be freed and the ice-crystal level is great and the amount of latent energy is large. Under such conditions the rising particles of air would have sufficient energy to reach well above the ice-crystal level, and at the same time the lower air particles would be accelerated upward for a long time, leading to high velocities and release a maximum of heat from condensation.—*R. G. S.*

Hail Forecasting

A recent study made by United Air Lines meteorologists indicates that in the central and eastern United States, *large hail* falls mostly in showers occurring between noon and 9:00 P.M. and especially between 3:00 and 6:00 P.M. Over half the hail falls were due to frontal activity, but insolational heating during the day apparently intensified the frontal "trigger action" to a great extent. The adiabatic charts and Rossby diagrams for hail days were studied but did not give very definite criteria for forecasting these hail storms. This was most likely due to the fact that the airplane soundings analyzed did not reach over 15,000 or 17,000 feet, and at those elevations the positive energy areas were usually either at their maximum width or still wide open, so that the real extent of the energy areas could not be taken into account. With the radiosondes now used this deficiency is eliminated. The following empirical results from the study, however, have a certain forecasting value:—

No falls of large-hail were reported in thunderstorms which occurred when the base of the conditional or convective instability region was above 7,000 feet. (This region began anywhere between 3,000 and 14,000 feet for hailless thunderstorms, but between 3,000 and 7,000 feet for storms with large hail.)

The proportion of thunderstorms producing large hail was one per 400 when the lapse rate (°F per 1,000 ft) was greater than 4.5°F in the convective or conditional instability region as indicated by the nearest sounding. The frequency was only one-third this great (one storm in 1,200) when this lapse rate was smaller than 4.5°F and the base of the instability level still below 7,000 ft. The lapse rate referred to was generally between 3.5° and 4.5°F for non-hail storms, but anything between 3.5° and 5.5° for storms with large hail.

Characteristic Properties of North American Air Masses[1]

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I. INTRODUCTION

In this discussion the term *air mass* is applied to an extensive portion of the earth's atmosphere which approximates horizontal homogeneity. The formation of an air mass in this sense takes place on the earth's surface wherever the atmosphere remains at rest over an extensive area of uniform surface properties for a sufficiently long time so that the properties of the atmosphere (vertical distribution of temperature and moisture) reach approximate equilibrium with respect to the surface beneath. Such a region on the earth's surface is referred to as a *source region* of air masses. As examples of source regions we might cite the uniformly snow and ice covered northern portion of the continent of North America in winter, or the uniformly warm waters of the Gulf of Mexico and the Caribbean Sea.

The concept of the air mass is of importance not in the source regions alone. Sooner or later a general movement of the air mass from the source region is certain to occur, as one of the large-scale air currents which we find continually moving across the synoptic charts. Because of the great extent of such currents and the conservatism of the air mass properties, it is usually easy to trace the movement of the air mass from day to day, while at the same time any modification of its properties by the new environment can be carefully noted.

Since this modification is not likely to be uniform throughout the entire air mass, it may to a certain degree destroy the horizontal homogeneity of the mass. However, the horizontal differences produced within an air mass in this manner are small and continuous in comparison to the abrupt and discontinuous transition zones, or *fronts*, which mark the boundaries between different air masses. Frontal discontinuities are intensified wherever there is found in the atmosphere a convergent movement of air masses of different properties.

Since the air masses from particular sources are found to possess, at any season, certain characteristic properties which undergo rather definite modification, depending upon the trajectory of the air mass after leaving its source region, the investigation of the characteristic properties of the principal air mass types can be of great assistance to the synoptic meteorologist and forecaster. We owe this modern analytical method of attack on the problems of synoptic meteorology and weather forecasting to the Norwegian School of Meteorologists, notably to J. Bjerknes and T. Bergeron. The analytical study of the synoptic weather map is based

[1]An abstract and revision of American Air Mass Properties, *Papers in Physical Oceanography and Meteorology,* published by Massachusetts Institute of Technology and Woods Hole Oceanographic Institution, Vol. 2, No. 2, 1933 (now out of print). A part appeared in the *Jn. Aeronaut. Sci.,* Vol. 1, No. 2, April, 1934, pp. 78-87. The revisions herein concern chiefly the Tropical air masses; but it should be understood that this discussion of the specific properties of the American air masses was a preliminary and pioneer work and the results have required some modification as further accumulations of aerological data and experience have been studied. A later study by Showalter is abstracted at the end of this article (pp. 109-113). The Bibliography lists many later papers which should be consulted by the serious student, as great advances are being made yearly.—*Ed.*

essentially on the identification and determination of the movement of air masses and fronts rather than of areas of high and low pressure as the entities of prime significance. The justification of this procedure is evident from the fact that the weather which is experienced in a given region does not depend upon the prevalence of high or low barometric pressure. Even in the same locality a high or a low may be accompanied by widely varying meteorological conditions. The weather in a given locality depends upon the properties of the air mass which is present or upon the interaction which is taking place between two air masses along a front in the near vicinity. The fundamental concept is that of the air mass, for upon the properties and movements of the individual air masses appearing on the map depend not only the weather in the area covered by each air mass but also the formation and intensification of fronts and the genesis, development and movement of lows or disturbances on the fronts. But in order that the full advantage of a careful analysis of the weather map may be utilized in weather forecasting, it is absolutely necessary that the thermodynamic properties of the air masses (lapse-rate and vertical moisture distribution) be approximately known. This is especially true of aviation forecasts, for just the meteorological elements which are of greatest interests to the pilot are those which are dependent upon the air mass properties. For good forecasting of convective turbulence, thunderstorms, horizontal visibility, fog and haze, cloud forms and ceiling, for just such forecasting full knowledge of the air mass properties is essential. Since it rarely happens that the forecaster has available for the current weather map aerological material sufficient for the satisfactory determination of the properties of the air masses present, an investigation of the characteristic properties of the typical American air masses should be of value both for practical weather forecasting and for a better understanding of the physical processes underlying our usual weather sequences. It was with this thought in mind that the paper here under consideration was written. The discussion was definitely restricted to a consideration of the homogeneous air masses individually, the investigation of the whole complicated frontal problem of converging air masses being left for later consideration in the light of more extensive American aerological data (see Bibliography).

II. CLASSIFICATION OF AIR MASSES

The study of synoptic weather maps indicates that air masses are entities having such definite characteristic properties that they may be classified and studied as distinct types. Since the characteristic properties of an air mass at any point depend primarily upon the nature of its source region and secondarily upon the modifications of the source properties which the air mass has undergone en route to the point of observation, any classification of air mass types must be based fundamentally on the air mass source regions, with perhaps a sub-classification based on later modifications of the source properties.

The air mass sources fall naturally into two groups, the tropical or subtropical, and the polar or sub-polar. The large areas on the earth of uniform surface conditions and comparatively light atmospheric movement lie almost entirely at high latitudes or at low latitudes. In middle latitudes, generally speaking, we find the zone of greatest atmospheric circulation or of most intense interaction between the warm and cold currents, i.e., air masses from the tropical and polar regions. Consequently, in middle latitudes the uniformity of conditions and the light air movement which must characterize a source region are generally lacking. Rather than the development of horizontally homogeneous air masses, we find here the rapid modification, in varied forms, with changing environment, of the characteristic polar and tropical air mass types. Thus the basis of any comprehensive air mass classification must be the distinction between the polar and the tropical source types, with a further distinction between the modified forms which these principal types acquire in middle latitudes during their later life history.

The air mass classification may be carried further by the sub-division of the polar and the tropical source types into continental and maritime groups, according as the source in each case is a continental or an oceanic region. Since the uniform source regions are always entirely continental or entirely maritime and since this is *the essential* difference between source regions in the same latitude, this distinction furnishes a satisfactory basis for a general grouping of the air masses from each latitudinal zone. Consequently, for the investigation of the properties of the air masses which may appear in a given locality, the most significant designation of the different individual polar and tropical air mass types is that based on the particular geographical area within which the air mass has its source. It is, of course, necessary to keep in mind the season of the year when considering the characteristics of any particular geographical source region, as these characteristics, especially in the case of the continental areas, change greatly from the cold to the warm season. It is also necessary to consider the modifications of the original source properties of the air mass types, effects which become more pronounced with the increasing movement of the air mass from the source region. Eventually the properties of the air mass become so fundamentally modified from the source properties that the mass must be given a special transitional designation.

Table II gives the complete classification of the principal North American air masses by geographical source regions, together with the principal transitional form for each air mass. The ordinary designation and the symbol entered for each air mass in the last two columns are those which appear on the Massachusetts Institute of Technology weather maps.

It is with the purpose of making this discussion intelligible to meteorologists who are not familiar with our local classification, and of making the American air mass data more directly comparable with the European data, that the general air mass classification outlined by Bergeron is also introduced in this paper.

In the general air mass classification which Bergeron[2] has suggested for climatological and comparative purposes, and which, at least in its broader features, Moese[3] and Schinze[4]

[2]T. Bergeron: Rechtlinien einer dynamischen Klimatologie, *Met. Zeit,* 1930, pp. 246-62.

[3]O. Moese und G. Schinge: Zur Analyse von Neubildungen, *Ann. d. Hyd.,* March, 1929, pp. 76-81.

[4]G. Schinze: Troposphärischen Luftmassen und vertikaler Temperaturgradient, *Beitr. z. Phys. d. fr. Atmos.,* Bd. 19, 1932, pp. 79-90; see also *Met. Zeit.,* May 1932, pp. 169-79.

TABLE II—CLASSIFICATION OF AMERICAN AIR MASSES

Source by Latitude	Source by Nature	Classification by Local Source Regions: Local Source Regions	Corresponding Air Mass Symbols (M. I. T.)	Season of Frequent Occurrence	General Classification, after Bergeron (International)
P	Continental	Alaska, Canada, and the Arctic	Pc	Entire year	cP or cPW, winter cPK, summer
		Modified in southern and central U. S.	Npc	Entire year	cPK
	Maritime	North Pacific Ocean	Pp	Entire year	mPK, winter mP or mPK, summer
		Modified in western and central U. S.	Npp	Entire year	cPW, winter cPK, summer
		Colder portions of the North Atlantic Ocean	Pa	Entire year	mPK, winter mPW, spring and summer
		Modified over warmer portions or the North Atlantic Ocean	Npa	Spring and summer	mPK
T	Continental	Southwestern U. S. and northern Mexico	Tc	Warmer half of year	cTK
		(Modified Form not found)	(Ntc)	?	
	Maritime	Gulf of Mexico and Caribbean Sea	Tg	Entire year	mTW, winter mTW or mTK, summer
		Modified in the U. S. or over the North Atlantic Ocean	Ntm (or Ntg)	Entire year	mTW
		Sargasso Sea (Middle Atlantic)	Ta	Entire year	mTW, winter mTW or mTW, summer
		Modified in the U. S. or over the North Atlantic Ocean	Ntm (or Nta)	Entire year	mTW
		Middle No. Pacific Ocean	Tp	Entire year	mTW, winter mTW or mTK, summer
		Modified in U. S. or over No. Pacific Ocean	Ntp or Ntm	Entire year	mTW
Upper Air (Sub-trop.)	Subsidence	Upper levels of the semi-permanent highs in the southern part of the Westerlies (Lat. 25°, winter; 40°, summer)	Ts or S	Entire year	cTW, winter
			Ts or S	Entire year	cTK, summer

follow in their discussions of European air masses, the essential distinction is still that between the tropical or the polar source of each air mass. However, Bergeron carries this zonal distinction one step further, distinguishing between real Arctic (A) and sub-Arctic, or Polar (P) air mass sources in the north, and between sub-Tropical (T) and real Equatorial (E) or Trade wind zone air mass sources in the south. Bergeron points out, however, that in northwestern Europe the Equatorial air masses play a negligible rôle, appearing only at high levels in the atmosphere, if at all. In the case of the North American air masses the distinction between Polar and Arctic air masses and that between Tropical and Equatorial air masses are both difficult to make and of little significance.

The principal source air masses in Bergeron's classification are the continental Arctic (cA), maritime Arctic (mA), continental Polar (cP), and so on through the mP, cT, mT, cE and mE groups.* Of course such a designation of air masses, while indicating very definitely the type of each air mass, is of necessity less precise in the information it gives to one thoroughly *familiar with the particular sources* in question than is the local classification by direct specification of the source of each individual air mass.

Air masses of Tropical and Polar origin are modified during their later history in either of two essentially different ways. If the air mass moves over a surface warmer than its own temperature at the ground, the tendency is then towards a warming of the lower strata of the air mass, i.e., an increasing thermal instability, and towards an increasing moisture content of the lower strata of the air mass, caused by evaporation from the warm surface. If, on the other hand, the air mass moves over a surface colder than its own temperature at the ground, the tendency is towards a cooling of the lower strata of the air mass, i.e., an increasing thermal stability, and towards a decreasing moisture content of the mass, caused by condensation from the cooled air strata. Evidently Polar air masses must normally undergo the first type of change when modified after leaving the source region, and Tropical air masses the second type. However, there may be exceptions in both cases, and any air mass may for a time be subjected first to the one and afterwards to the other type of influence.

In Bergeron's general air mass classification, modification of the source properties of the air mass, which in the local classification is indicated by the N (transitional) group, is indicated by a W (warm) or a K (cold, *kalt*) distinction according as the recent modification of the air mass has been of the second or the first type mentioned above. The warm (W) designation indicates that the air mass is warm relative to the surface it is moving over, the cold (K) designation indicates that it is cold relative to the surface it is moving over. Thus in the general classification of air masses the source designations cP, mP, cT, and so forth, when applied to air masses which have left their source regions, appear in the modified forms cPW (continental Polar Warm), cPK (continental Polar Cold), mPW (maritime Polar Warm) and so forth, depending upon the type of modification which the air mass has undergone during its recent history. It should be stressed that this warm and cold designation has nothing to do with the evidence by the air mass of

*This manner of designation was introduced into the practice of the U. S. Weather Bureau in 1939; it has been in extensive use abroad and become understood internationally.—*Ed.*

a high or a low temperature, but only as to the evidence of a temperature near the ground higher or lower than that of the surface beneath. This warm and cold distinction is not always easy to make, as the passage of the air mass from ocean to continent or the transition from day to night may reverse the sign of the difference of the air temperature from that of the surface beneath. In the present discussion the policy will be to consider only the general tendency in the change of properties from one day to the next in the history of the air mass when determining the warm or cold designation. Continued or increasing surface stability from day to day indicates a warm air mass (W), continued or increasing instability from day to day a cold air mass (K). This thermodynamic classification of air masses into warm and cold groups is essentially differential in nature, depending as it does upon changes produced in the air mass properties by boundary surface-temperature differences. In contrast to the significance of the source classification which depends upon the conservatism of certain of the air mass properties, the significance of the W and K classification lies in the modification of the non-conservative air mass properties.

There are a number of conditions which are more or less frequently met with in the synoptic study of air masses which may locally or temporarily render difficult the proper classification of an air mass. In particular the disturbing effect on the air mass properties of the surface over which the mass is moving and the consequent formation of a ground layer with its own peculiar properties cannot be overlooked. When this influence is regular and continuous, it gradually affects the entire mass until it becomes characteristic for the properties of the mass. On just such influences depend the thermodynamic "warm" and "cold" classification of air masses already mentioned. But mechanical turbulence produced by surface irregularities, and the radiational effects of a single night and insolational heating of a single day often produce ground layers with properties which may be neither permanent nor characteristic of the air mass, and which consequently must be allowed for in the discussion of the air mass properties in particular cases. Especially troublesome are the large radiational-insolational effects at the surface in dry continental air masses during the warm season. In the central U. S. this diurnal surface temperature change may amount to more than 20C°, may produce an afternoon unstable layer more than 2 km thick, and may change the air mass from the warm to the cold type. Föhn and subsidence effects are also frequently to be noted, but they are usually definitely characteristic of certain air masses under certain conditions, and belong as such to the characteristic air mass properties, not being to any great extent diurnally variable.

III. SIGNIFICANCE OF THE PROPERTIES OF THE PRINCIPAL AIR MASS TYPES IN WINTER

It is impossible in a short review to attempt a full summary of the seasonal properties of all of the air masses listed in Table II. Consequently, this discussion will emphasize the air mass types, Pc, Pp, and Tg, which are dominant in determining our winter weather in the United States. It is during the winter season that the air mass contrasts become most significant. The following discussion will illustrate the applicability of the analytical method to weather forecasting, without pretending to indicate more than a small fraction of the possibilities met with in actual practice.

In Table III we find tabulated for each of the three principal air masses at two different aerological stations the mean values of the temperature, T, the specific humidity,* w, the relative humidity, *R.H.*, and the equivalent-potential temperature, θ_E, at the ground and at the successive km levels above sea level. These mean values are simply the averages of a number of ascents chosen as typical of each air mass type at each station. They represented the best observational evaluation which could be made of the so-called characteristic properties of each air mass type at the time this study was undertaken.

[Since then a further analysis by Mr. Showalter of the 1935-36 aerological material has been published (*Mon. Wea. Rev.*, July 1939) which deserves careful study. At the end of this article some excerpts from his paper are reprinted, but the original gives a series of tables and cross-sections of mean values and frequency diagrams for each air mass, which show that there is generally a large variation from case to case of the same air mass type even in the same season, so that mean values must not be accepted too literally. They do not necessarily give a *typical* picture because so many features which probably never *all* occur in any *one* sounding are averaged together. But the average picture is necessary and valuable as an orientation for the student, forecaster, and researcher.—*R. G. S.*]

An explanation of the full significance of θ_E cannot be made here[5] but it may be stated in a general way that turbulent mixing of an air stratum tends to effect isothermalcy of θ_E, whereas a rapid vertical change in θ_E indicates marked atmospheric stratification. If θ_E increases with elevation, the possibility of thermal convection in the atmosphere is practically excluded, whereas if θ_E decreases with elevation, the atmosphere is potentially unstable, an instability which becomes actual with sufficient vertical displacement of the affected stratum. Such is the displacement which may occur at a warm front. The amount of displacement necessary to effect actual instability of the affected stratum is less, the higher its relative humidity.

In Table III the first station included for each air mass type is the one which gives the best indication of the characteristic properties of the air mass as it advances directly, fresh from the source region. The second station is chosen to indicate the most important modified form in which the air mass appears during its later history. Thus for the Pc air mass, Ellendale indicates the characteristic cold wave type in the middle west, while Boston indicates the rather fundamentally modified cold wave type in the northeastern United States. For the Pp air mass, Seattle indicates the characteristic properties as the air mass approaches the northwest coast of the United States fresh from the source, while Ellendale indicates the importantly modified form which the air mass assumes by the time it

*The quantity w is not quite the specific humidity, defined by $q=0.622e/p$ but the mass ratio of water vapor to dry air, defined approximately by $w=0.622e/(p-e)$ where e and p are water vapor and total atmospheric pressures. (See, Namias, article III.).

[5]See C.-G. Rossby: Thermodynamics Applied to Air Mass Analysis, *M. I. T. Meteorological Papers*, Vol. 1, No. 3; also Articles III and IV of Mr. Namias' series, in this booklet.

TABLE III. Pc, Pp, AND Tg AIR MASSES—WINTER

Air Mass	Station	Element	Elevation Above Sea Level (km): Surface	1	2	3	4
Pc	Ellendale	T (°C)	—26.1	—25.3	—20.1	—21.5	—25.4
		w (g)	0.32	0.35	0.60	0.50	0.45
		$R.H.$ (%)	82	80			
		θ_E (°A)	250	256	272	280	288
	Boston	T (°C)	—6.3	—14.3	—18.0	—23.0	—29.0
		w (g)	0.9	0.6	0.5	0.3	0.2
		$R.H.$ (%)	43	55			
		θ_E (°A)	267.0	268.0	274.3	279.0	283.3
Pp	Seattle	T (°C)	8.3	0.0	—8.3	—14.3	—19.3
		w (g)	4.4	2.7	1.5	0.8	0.4
		$R.H.$ (%)	75	88			
		θ_E (°A)	292	289	288	289	294
(Npp)	Ellendale	T (°C)	—1.2	7.0	0.8	—6.5	—14.0
		w (g)	3.0	3.0	2.2	1.5	1.1
		$R.H.$ (%)	84	40			
		θ_E (°A)	284	299	300	301	302
Tg	Groesbeck (includes much T_s above 1 km)	T (°C)	18.8	14.0	13.0	7.5	
		w (g)	12.6	10.4	4.1	1.2	
		$R.H.$ (%)	90	95	40	15	
		θ_E (°A)	327	326	318	312	
	Boston	T (°C)	14.0	13.7	8.7	2.0	—3.7
		w (g)	8.8	6.5	6.2	4.6	2.9
		$R.H.$ (%)	82	60	86	84	85
		θ_E (°A)	310	314	319	318	319

reaches the interior of the continent, the characteristic NPP air mass type. Finally, for the TG air mass, Groesbeck indicates the characteristic properties as the air mass first advances northward from the Gulf of Mexico, while Boston indicates the modified form which the air mass assumes as it moves northeastward, approximating the characteristic NTM (NTG) type.

[For a critique indicating some limitations involved in analyzing the properties and movements of air masses solely by comparison of the T, q, and θ_E values at standard elevations in the free air as done here, see the paper by Prof. Rossby in *Bull. Amer. Met. Soc.*, June-July, 1937, pp. 201-9. Isentropic analysis was developed to avoid these limitations (see article X in this booklet).—*Ed.*]

IV. WINTER AIR MASS PROPERTIES

The Pc Air Masses. Let us consider now the significance of the characteristic properties of each of the air mass types tabulated in TABLES III-V, bearing in mind that the entire discussion applies essentially only to the winter months. The PC air masses are those which originate over the snow and ice covered regions extending from the interior of Canada northwestward over Alaska and northward into the Arctic. The passage of an air mass of this type over Ellendale marks the advance of the typical cold wave from the Canadian northwest, the characteristic properties of the air mass observed at Ellendale being practically those acquired in the source region. We note at once from Table III the extreme coldness of this air mass at Ellendale but also the fact that the temperature increases above the ground, reaching a maximum at about the 2½ km level. Actually we find that in the early stages of the cold air outbreak when the wind is still moderate to fresh northwest or north from the ground up to high levels, the temperature decreases through the first few hundred meters above the ground. This is a result of the mixing affected by mechanical turbulence. Consequently a minimum temperature is often found just below the 1 km level, with a marked inversion above. In the later stages of the cold air outbreak, as the wind diminishes and becomes very light at the ground, radiational cooling effects a lowering of the air temperature at the ground, while subsidence effects a raising of the temperature at upper levels.* Thus, with the possible exception of a shallow turbulence layer near the ground, the PC air mass shows very marked stability through the lower 2 km, usually even a rather large temperature inversion, which tends to increase with the aging of the air mass over the Plains states and the Mississippi Valley, at least as long as the mass remains over a snow or ice-covered surface. Passage of the PC air mass during its progress southward over bare ground or especially over open water, leads to a rapid heating and moistening of the lower strata. Consequently the typical NPC air mass in more southerly latitudes is likely to be characterized by rapidly diminishing stability, but during the winter

*See J. Namias: Subsidence in the Atmosphere, *Harvard Met. Studies* No. 2, 1934.

season it never becomes really unstable in the Mississippi Valley before it reaches the Gulf of Mexico.

We notice from Table III that the specific humidity at Ellendale is very low at all levels but with a pronounced maximum at the 2 km level, whereas the relative humidity is uniformly rather high at lower levels. The indication is that the radiational cooling of the surface air strata to very low temperatures has led to condensation and deposit of moisture from these strata. Generally speaking, values of specific humidity less than 1 g and a rapid increase of θ_E with elevation, are characteristic of the Pc air masses.

A Rossby diagram for Pc air (cPW) at Ellendale shows the typical extremely characteristic form of the θ-w curve for this type of air at its source. And this type form is characteristically conservative at least as long as the air mass remains over a snow-covered surface. The curve is nearly vertical, showing a small increase of w at intermediate levels, but with w values less than a gram at all levels. The potential and equivalent-potential temperatures increase some 40C° in 4 km; consequently the existence of any mechanically produced turbulence effects rapid transfer of heat downward in the turbulent layer.

The normal flying conditions prevailing in the Pc air mass in winter in the northwestern and central United States may be readily surmised from the air mass properties. The condition will be one of extreme cold at the ground, but the temperature will be much higher at intermediate levels (1 to 2 km above ground). Bumpiness will be markedly absent, flying being unusually smooth, except perhaps for the shallow turbulence layer near the ground when the winds are fresh. Condensation forms are characteristically absent, though at the beginning of a Pc outbreak a low stratus or stratocumulus cloud deck may mark the top of the turbulence layer, but is likely to dissolve rapidly. Otherwise the air mass should be cloudless, high clouds indicating the presence of another air mass aloft. Occasionally the radiational cooling of the lower strata of the air mass leads to the local formation of frost smoke or ice crystal fog at very low temperatures. The intensification and lowering of the subsidence inversion and the disappearance of all surface wind in the stagnant and aging Pc air mass favors the accumulation of smoke and dust pollution in the lower levels of the atmosphere. Consequently the Pc air mass during its later history, and especially in industrial regions, is likely to be marked by unusually dense smoke haze below the inversion and by sharply marked haze layers. But the air mass in the central United States is usually too dry for the formation of any extensive dense fog below the subsidence inversion. Fog is likely to appear in the old Pc air mass first upon the approach of a warm front with a lowering warm front cloud deck and precipitation. This is a very dangerous condition for flying, as it favors not only fog formation but also heavy ice formation on the plane. However, this condition pertains essentially to the warm front zone and not to the Pc air mass as such.

In the northeastern United States, and generally along the Atlantic seaboard, the Pc air mass properties vary significantly from the characteristics which they show in the Mississippi Valley and the northwest. This is probably to be attributed to the following facts:

(1) The existence of considerable

areas of open water in the Great Lakes in winter. The rapid heating and moistening which this effects in the lower strata of the cold, dry, stable Pc air masses are evidenced by the persistent low cloudiness, convectional snow flurries and relative warmth of the onshore Pc current which has crossed any of the lakes. Presumably this instability becomes rapidly less at upper levels.

(2) The presence of the highlands and numerous ridges of the Appalachians which must be crossed in the eastward advance of the Pc current. On the western slopes of the Appalachians the effect of the forced ascent of the Pc current which has already been rendered conditionally unstable over the Great Lakes is to cause general cloudiness and more or less continuous snow flurries as long as the cold air flow continues unabated. On the coastal slopes the föhn action causes a partial or total dissipation of the cloud deck and snow flurries. However, the rough terrain maintains a steep lapse-rate through the lower km or more, with low relative humidity and relatively high temperature at the ground.

(3) The increasing depth and strength of the Pc outflow in the eastern United States. The rapid intensification and increasing vertical depth characteristic of cyclonic disturbances moving northeastward over the United States causes an acceleration of the Pc outflow over the eastern United States especially at higher levels. This helps to increase the turbulence in this air flow at low levels, while the rapid transport southward of air from Arctic sources at upper levels maintains a good lapse-rate aloft.

These effects are to be seen clearly in the data for Boston in Table III. We note the high surface temperature in contrast to that at Ellendale, and the steep turbulence lapse-rate in the first km and the moderate rate above, such that at the 3 and 4 km levels the characteristic Pc temperatures are notably colder at Boston than at Ellendale. We observe also the low surface relative humidity increasing with elevation and the almost constant value of θ_E in the first km, both of which indicate thorough turbulent mixing, while even at higher levels θ_E increases much more slowly than in the Pc air at Ellendale.

These characteristics of the Pc air mass in the northeastern United States and along the Atlantic coast explain the typical flying conditions which are met with during its prevalence in this region. As long as the cold air flow continues actively, the roughest and gustiest flying conditions which may be experienced in winter are to be expected up to considerable elevations. Along the Atlantic seaboard the skies are likely to be almost clear, with excellent visibility and, normally, only scattered fracto-cumulus or cumulus clouds, though scattered light snow flurries sometimes occur in the afternoon. In the Appalachians and on their western slopes one may expect a rather low broken cloud deck, often continuous over considerable areas, with frequent, extensive and often heavy snow squalls. After the force of the Pc outbreak has been spent, so that the air mass becomes relatively stagnant, then subsidence, radiation and the cessation of the turbulent mixing all act to establish quickly in the Pc air mass the low inversion, thermal stability and dense stratified smoke haze so characteristic of this air mass in the middle west. Of course, this change is accompanied by the complete disappearance of the instability snow flurries, cloudiness, and gustiness.

There is nothing very distinctive about the synoptic situation which leads to the outflow southward of Pc air from the source region in North America. Usually the outbreak is preceded by a gradual building up of high pressure and low temperatures in the source region, often without any very obvious atmospheric movements. Sometimes the cold air outbreak follows quickly on the first appearance of the cold anticyclonic circulation in the north, but often there is an extended delay, during which cold air surges in the north seem on the point of breaking out, only to recede again. Sometimes the final outbreak follows on the passage inland from the Pacific Ocean of a disturbance usually already occluded in the Aleutian Island region. Other times the cold air suddenly advances without any very obvious disturbance to start it moving. Usually, however, when this happens, one or more disturbances are likely to form along the cold front as it advances southward. This is especially likely to happen as the cold air pushes southward to a point where it begins to come in contact with warm moist air from the south, particularly Tm air from the Gulf (Tg). The disturbances usually appear first as flat wave formations on the cold front, but are likely to develop rapidly into intense occluding cyclones which greatly accelerate the movement of the Pc air behind them to more southerly latitudes. In such a development the advance of the Pc air mass and cold front is likely to be checked at first by the warm air current. Then very quickly an extensive overrunning of the Pc air by the Tg current takes place. The overrunning circulation tends to develop cyclonically or counter clockwise, with heaviest precipitation extending far back of the cold front. As J. Bjerknes[6] has pointed out, this type of overrunning is to be expected only when the warm current is unstable; but as we shall see presently, marked conditional instability is always characteristic of our Tropical air from the Gulf. Refsdal[7] was the first to call attention to the importance of the rôle which conditional instability in the atmosphere may play in the development of frontal disturbances and in the regeneration of old occluded disturbances. The writer is inclined to the belief that the moist-conditional instability (Feuchtlabilität) of our Tg air masses is a very important factor in the control of the interaction between the warm and cold air mass and the stability or the tendency to occlusion shown by wave disturbances on the cold front between such air masses. An excellent instance of the weather sequence with the gradual advance of a cold front followed by an extensive Pc air mass may be found on the weather maps and upper-air data for Jan. 13-16, 1930 in the U. S. Following the first sweep of the cold air southeastward to the Gulf of Mexico, the whole frontal system followed by the Pc outflow advances steadily eastward until the whole eastern and central U. S. is covered by the cold air. On the other hand, it sometimes happens that the cold air outflow from the source region takes place in one sudden abrupt outbreak, without the development of any well-marked disturbance. In this case the cold air sweeps everything before it, apparently too rapidly for any localized disturbance to develop on the cold front. Furthermore, these very fast moving cold air outflows usually occur only on the heels

[6]Exploration de quelques perturbations atmosphériques. . . . *Geofys. Publ.*, Vol. IX, No. 9.

of a preceding outbreak which has as it were paved the way for it by displacing the warm maritime air masses, so that we find the fresh Pc mass displacing only moderately cold dry continental Polar air masses (Npc) as it moves southward towards the Gulf. At this point it is necessary to inquire a little further as to what happens to the original Pc, or cPW, air-mass properties in the course of the later history of the mass.

Modification of the Winter Source Properties of the Pc Air to Npc Air

Following the first outflow of the Pc mass from the source region, where its properties are essentially those shown by the Ellendale data, there are three principal influences which may be at work to change these properties. Observed changes of the properties are to be explained then in terms of one or more of the following modifying influences:

(1) Supply of heat by contact with and radiation from the surface beneath, and supply of moisture by evaporation from this surface. This influence tends to make the lower strata of the Pc mass increasingly warm and moist, or to change its properties from cPW to cPK.

(2) The effect of subsidence, which tends to intensify the temperature inversion at a low level and to effect a general warming of all the upper strata.*

(3) The effect of turbulence occasioned by a rough undersurface in developing a thoroughly mixed layer in hilly or mountainous regions.

As the cold Pc air mass moves southeastward from the Canadian border west of the Great Lakes across the Plains states and down through the Mississippi Valley towards the Gulf of Mexico, the modifications of its properties depend principally upon the extent of the snow cover, for on this depends the effectiveness of the first and most important of the above three influences. The path of the cold air advance is over flat country all the way to the Gulf of Mexico, so that turbulence effects become no more pronounced than they are at Ellendale.

TABLE IV.

Pc AIR—WINTER (cPW BECOMING cPK)
(Surface and first 3 km at)

Royal Center, Ind.				Broken Arrow, Okla.			
T °C	w g	RH %	θ_E °A	T °C	w g	RH %	θ_E °A
—22.5	0.45	91	251.5	—14.8	0.95	86	260.5
—19.5	0.48		262.0	— 8.8	1.15		275.3
—15.8	0.48		276.5	— 8.0	1.20		285.7
—18.0	0.85		286.0	—10.3	1.03		294.3

As long as the air mass continues moving over snow cover, there is only a very slight warming and increase of moisture to be observed. The air mass retains the characteristic coldness of the surface stratum, though the cold layer becomes shallower, and the temperature inversion sharper. This is doubtless due to subsidence, i.e., the sinking and spreading of the cold air mass. The Pc values of the air-mass properties given in Table IV for Royal Center, Ind., and Broken Arrow, Okla., are averages of ascents for the same period as those at Ellendale, January, 1930, in cases of minimum modification of the original Pc properties, i.e., in cases of the maximum coldness and dryness of the original Pc air. Such cases are those where the trajectory of the air mass the entire distance to the point of observation has been over snow-covered ground. And this will be the case only for the rapid outbreaks of cold air where the freshly outflowing cold air is displacing only slightly less cold air of the same source. Otherwise the modification of the Pc properties takes place much more rapidly.

*See J. Namias: Subsidence in the Atmosphere, *Harvard Met. Studies* No. 2, 1934. Also see figs 18, 19, on pp. 70-71 of this booklet.

We notice in passing southward from Ellendale that the surface temperature of the Pc air becomes somewhat warmer, as would be expected, but remains extremely cold. At Royal Center and Broken Arrow the surface relative humidity remains very high, and the temperature inversion increases slightly and becomes lower in elevation. w still increases with elevation at all stations. These changes are just what we should expect from a moderate degree of subsidence and a slight increase of heating by terrestrial radiation, and as a result of evaporation from a snow surface which is not here so cold as it is farther north in the source region. Especially during the daytime is the sun much more effective at the lower latitudes in raising the temperature of the snow surface, and doubtless more heat is transmitted from the ground beneath, which is warmer, through a snow cover which is thinner.

The characteristic curves for these stations representing their average Pc properties on the Rossby diagram are strikingly similar. And equally uniform is the clear, dry, cloudless weather attending the passage of the Pc air mass, as long as it retains this characteristic temperature and moisture distribution. In the later stages of the Pc outbreak, if the cold air outflow weakens and the conditions become stagnant, the gradually lowering and intensifying subsidence inversion is naturally attended by markedly decreased visibilities below the inversion, particularly in industrial regions where smoke pollution of the atmosphere is pronounced. It is, however, the exception rather than the rule for such stagnation to follow in the case of a strong outbreak of Pc air. The stagnant anticyclonic condition with low subsidence inversion and poor visibility usually occurs in the U. S. in the southeastern states, and only with mild outbreaks of Pc air. And it seldom leads to fog formation, because the cold air, being continental, is usually too dry. In Europe, on the other hand, where the stagnation of the cold air masses is more frequent and prolonged, and where the Polar and Arctic air masses are frequently of maritime origin and comparatively moist, this condition often leads to the widespread formation of fog or low stratus cloud, with practically zero visibility below the temperature inversion. Above the inversion visibilities are always excellent. Of course, active overrunning of the Pc mass by a warmer moister air current completely changes the type of weather. Such overrunning by warm Pacific air in the Rocky Mountain region or by warm Gulf air in the Plains states produces the typical blizzard conditions for these respective regions.

As soon as a typical Pc air mass leaves snow-covered territory, the modification of its characteristic properties becomes very rapid. Heating of the lower strata by contact with and radiation from the ground, and increase of moisture by evaporation from the ground surface, effect a rapid transformation of its properties, most marked near the ground and diminishing aloft. At the same time that the lower strata of the air mass are being rapidly warmed and moistened from the ground, there is doubtless also some direct radiational heating of the upper strata in response to the increased terrestrial radiation from the warmer ground surface. The initial heating of the upper strata does not have to wait on the establishment of thermal convection. The ground heating of the air mass is especially rapid during the daytime in southerly latitudes, where the clear skies characteristic of Pc weather favor marked insolational heating of the ground, and in particular when the Pc mass is displacing a Tropical maritime air current, in which case the ground will be particularly warm and moist and the cold air mixes with remnants of the warm moist air. In this way under ordinary conditions the southern portion of an advancing Pc mass becomes rapidly transformed, as it moves southeastward across the U. S., from the typical source cPW properties of the air mass to a cPK, or unstable air mass, in which the water vapor falls off rapidly with elevation to the low values aloft initially characteristic of the Pc mass. This modified cPK form of the Pc mass is designated as NPC on the M. I. T. maps.

In contrast to the increasing instability of the Pc air mass as it is transformed to NPC, should be noted the increasing stability of the PP air mass as it is transformed to NPP. In the extreme case the NPC mass, or part of it, may be turned northward again, become a return flow over a colder surface, and tend to a reëstablishment of cPW stability. In general, this rapid modification of the properties of the southern portion of the Pc air mass relative to the northern portion means that the initial horizontal homogeneity of the air mass is destroyed, or, in other words, the isentropic surfaces do not remain in their initial horizontal position. This fact has long since been remarked by Bergeron[8] as characterizing all Polar air masses, and offered as the explanation of the readiness of the Polar air masses to form secondary fronts within themselves. Any wind discontinuity quickly produces a discontinuity of the other elements under these conditions.

Generally speaking, the transformation over land of the cPW to cPK air mass has little significance for the general weather conditions. Visibility at the ground is improved, and apart from the possibility of the formation of a few cumulus clouds during the day, the skies remain clear. But the effect of open water on the cPW air mass in winter is immediate and pronounced, the mass becomes cPK forthwith. The addition of moisture and heat to the cold dry air mass takes place very rapidly from any water surface. In surprisingly quick time the lower strata of the air mass are rendered convectively unstable, and heat and moisture are carried to successively higher levels by active convection. The quantitative effect of this influence on the vertical distribution of temperature and specific humidity will of course depend upon the Pc air mass properties, the warmth of the water surface, and the length of the sojourn of the cold air over the open water. The extreme degree to which such a modification may go is indicated by a comparison of the properties of our PM air masses at Seattle with our Pc masses at Ellendale. Originally these two air masses have about the same properties, but the PP is modified during its passage over the ocean. A still more extreme case is that of our Pc air which may reach the Gulf of Mexico, under favorable winter conditions, still as a very cold, dry and essentially stable mass. But the winter northers in the Central American countries, which are invariably continuations of our Pc outbreaks across the Gulf of Mexico southward into the sub-Tropical circulation, are typically only moderately cool at the surface and attended by

[8] T. Bergeron und G. Swoboda: Wellen und Wirbel an einer quasi-stationärer Grenzflache über Europa, *Veröff Geophys. Inst. Univ. Leipzig*, Ser. 2, Bd. III, no. 1, 1924.

extremely heavy convective precipitation. This change which takes place in only 36 to 48 hours over the warm waters of the Gulf amounts often to a warming of the surface stratum of the Pc mass by 25° or even 30C°, and an increase of w from perhaps 1 to about 15 g. Unfortunately few soundings from these regions are available, but they indicate a rapid approach to if not actual convective equilibrium under these conditions.

For the north-central and northeastern U. S. the modification of the Pc air mass properties by the Great Lakes is very important. During the late fall and early winter, while the Lakes are still free of ice, this effect is particularly well marked. On the coldest Pc outflows with a northwest wind blowing across Lake Michigan, for example, which is less than a hundred miles wide, the temperature may be 10°C, or even more, higher at Ludington on the east shore than at Green Bay on the west shore. And whereas the sky will be cloudless at Green Bay, at Ludington there will be low ceiling and constant snow flurries. And this difference is characteristic of all the Lake stations; passage of the cold air across the Lakes produces always a low Nbst or Cunb cloud with almost continuous snow flurries. Apparently the instability layer produced over the Lakes is rather shallow, as evidenced by the characteristic lightness of the precipitation and the amount of the temperature increase in the ground stratum of the air mass. However, it is marked enough so that considerable cloudiness with snow flurries characterizes the Pc current from the Lake region eastward up the moderate slopes to the Appalachian Divide. The surface values of w are found to have increased in passing over the Lakes from a normal value of about 0.5 g to perhaps 2.0 g, and the temperature to have been raised by some 10C°. This modification is a maximum at the ground, disappearing at some 2½ km elevation. In fact, the thermal influence of the Great Lakes favors very markedly the formation of slowly moving shallow cold fronts in this region, which are indicated at the ground by a well-marked wind-shift from southwest to northwest and a corresponding trough in the isobars. But observations of winds aloft show always that at as low as 2 km elevation this discontinuity has completely disappeared, the wind at that level being uniformly northwest. This means that from the Lake Region eastward to the Appalachians the tendency is to a continuous overrunning of the somewhat modified Pc air at the ground by colder Pc air aloft coming more directly from the source. This doubtless favors the appearance of the instability snow flurries which characterize the Pc weather in this entire region.

The Pp Air Masses. The Pp air masses are those originating in the Arctic or sub-Arctic, which reach the Pacific coast of the United States or Canada after a more or less prolonged sojourn over the waters of the north Pacific ocean. Initially these air masses come from the Arctic source regions with properties which probably approximate those which we have found to be characteristic of Pc air, but prolonged heating and moistening over the warm waters of the north Pacific produce, to an extreme degree, the type of modification which we found to be effected in the Pc air masses in winter by the Great Lakes. They are changed from a cold, dry, extremely stable condition to one of marked conditional instability with a comparatively high moisture content in the lower strata.

The data from Seattle in Table III give the average conditions in the Pp air mass as it arrives on the north Pacific coast direct from more northerly latitudes. When these air masses have made a swing southward and reach the Pacific coast from the west or west-southwest they are still warmer and moister than indicated by the data from Seattle in Table III but especially so in the upper strata. Consequently, the convective instability is less pronounced and is displaced upward to the higher strata of the air mass.

If we compare the properties of the Pp air mass at Seattle with those of the Pc at Ellendale, we note that, whereas at the ground the former is 34°C warmer than the latter and has a specific humidity nearly 15 times as great, at the 3 and 4 km levels the temperature difference is only 6 or 7°C, while the difference in the specific humidity almost vanishes. Probably 3 km is about the upper limit to which convection has penetrated in the Pp air mass. The equivalent-potential temperature at this level is somewhat lower than at the ground but it begins to increase rapidly above.

As would be expected from the conditional instability and high moisture content of the lower strata of the Pp air mass on the Pacific coast, flying weather conditions are not favorable in the Pp current. Visibility is good in the absence of condensation forms but we find normally a rather low broken cloud deck, which may be solid over large areas, with extensive cumulo-nimbus formations and convective showers which are frequently rather heavy and may follow in quick succession. Winds are likely to be gusty, and flying extremely bumpy. On the western slopes of the coastal mountain ranges the clouds are likely to be low and thick and the precipitation (snow in the mountains) to be heavy and almost continuous. Such conditions will persist as long as the steady onshore flow of the Pp air continues.

It happens quite frequently in winter that a westerly flow of air from the Pacific (usually Pp air masses) continues for considerable periods of time, so that the weather conditions prevailing over the greater part of the United States may be dominated by these air masses. We have just seen the unfavorable weather conditions which prevail in the Pp air current in the Pacific coast region but this air mass is importantly modified in crossing the western mountain ranges to the interior of the country in the following ways:

(1) Much of the moisture contained in the lower strata of the Pp air mass is condensed in heavy convective showers and the heat of condensation is supplied to the air mass strata at intermediate levels.

(2) The descent of the eastern slope of the Rockies dissipates the cloud deck and convective showers in the Pp current, while the heat of condensation already supplied to the intermediate strata of the air mass and adiabatic compression (föhn effect) cause a marked warming of the air mass at the 1 km level and above.

(3) The warming of the surface strata is checked by radiational and contact cooling over the cold continental surface and perhaps also by mixing with remnants of the cold Pc air.

The effect of these various influences acting on the Pp air mass is readily seen by comparing the Pp data in Table III from Seattle with those from Ellendale, which may be taken to typify inland conditions in this air mass. At Ellendale the Pp air mass shows a very large temperature inversion and constant value of w in the first km, instead of the instability found at Seattle. The temperature increase over the Seattle values shows a maximum at the 2 km level. Probably, however, the marked increase in T and w found at 1 km and above, in passing from Seattle to Ellendale, is considerably in excess of the change which actually occurs in the air current itself, for the Pp current which reaches Ellendale comes from the west or west-southwest. It is not one of the direct northwest currents such as Seattle's data represent but one of the westerly currents which is doubtless appreciably warmer and moister at upper levels than the fresh Pp air from the northwest. However, the Ellendale data are absolutely typical for the Pp air masses which have crossed the mountains to the interior of the United States. The modification from the typical coastal Pp properties is so pronounced that the symbol Npp instead of Pp is used for these air masses in the central United States.

The weather conditions which prevail in the Npp air masses east of the Rockies are in general the best which are experienced in winter. Air movement is usually moderate, and convective turbulence is experienced only in the afternoon when the large diurnal temperature range found under these conditions may be sufficient to overcome the normally prevailing surface temperature inversion. Temperatures are notably mild, skies clear except possibly for some scattered high clouds and visibilities generally very good except for the possible occurrence of morning smoke haze in industrial regions.

When, however, an advancing Pp air current comes in contact with a strong outflow of cold Pc air on the eastern side of the Rockies, it is likely, as the warmer current, to ascend and advance over the Pc current along a typical warm front. Such a development may cause severe blizzard conditions in the Plains states, but this is, of course, a frontal phenomenon and in no sense a characteristic of the Pp air mass.

Polar Atlantic Air Masses in Winter

[Owing to the prevailing west-east air movement the North Atlantic is not an important source region for air masses affecting North America, especially in winter when this movement is strongest. The source region for Pa air affecting the U. S. is the part of the N. Atlantic adjacent to the continent and north of the warm Gulf Stream, from the Gulf of Maine north, where the waters are abnormally cold the year round; in winter these waters only effect a modification of Pc air masses which move off the continent and which may occasionally retrograde westward as Pa air onto the eastern coast, where its properties are distinct from those of other air masses. As it moves off the coast the anticyclonic circulation of a large body of Pc air brings the old Pc, modified into Pa, onshore again as a NE wind backing to SW, which return flow may be strengthened by the approach of a disturbance (warm current) from the S or SW; it brings a sudden change of weather but seldom reaches beyond the Appalachians.

The vertical structure of Pa on the north Atlantic coast of U. S. (Boston, e.g.,) is characteristically dry and stable aloft, with a fairly shallow, cold and moderately unstable moist layer at the surface; the lower levels are colder, drier and more stable than in PP air (at Seattle, e.g.) because the waters (0°C) are colder and the air has spent much less time thereover. The upper levels show the typical structure of old PC air, much warmed and dried by subsidence, and above 3 km the warm-front inversion (often due to tropical air) of the next approaching disturbance may already be present.

On the coast, therefore, the PA air has a surface temperature near freezing or somewhat below, with a good lapse-rate through the first km, more or less, the relative humidity rather high ($w = 2$ or 3 g/kg) and increasing to saturation or nearly so at the top of the convective layer, where there is a deck of thickening Stcu; but Cunb and showers are prevented by the marked inversion a short distance above the cloud base. Above the inversion it is dry, and clear unless there are warm-front clouds still higher. At Boston, for example, the PA air arrives rather suddenly following a rapid backing of the wind from NW to NE, with thickening low Stcu clouds scudding in from the E; later light flurries or fine mist begin, making poor the already failing visibility, and the temperature has risen perhaps as much as 10°C; the front of the maritime air works steadily inland, abruptly terminating cold weather, till it nestles in the lee of the eastern ridges of the Appalachians. Farther south this type of onshore flow is much warmer owing to the nearby Gulf Stream, and unstable to higher levels due to its longer maritime history; but here reports are often insufficient to prove Polar origin—it should be called PA or NPA if traced from PC, otherwise TA or NTA. NPA is PA air warmed and unstable beyond typical PA conditions.—*R. G. Stone.*]

The Tropical Air Masses. The TG air masses are those which originate over the warm waters of the Gulf of Mexico and Caribbean Sea, or further southeast in the Tropics. Scarcely to be distinguished from them in characteristic properties are the TA air masses from the Sargasso Sea region, so that this discussion really applies to both. The TP air masses are the Pacific counterparts of TG or TA. The transitional forms of these masses are usually referred to simply as NTM (transitional Tropical maritime). There is no attempt to make the very difficult distinction between NTG and NTA.

The Tropical Pacific Air Masses

The source of the TP air masses is that portion of the Pacific Ocean lying west and southwest of southern California, roughly between latitudes 25° and 35°N. During the cold season this region is one of markedly steady weather conditions, as it is normally the winter seat of the stationary Pacific anticyclone. Over the Pacific Ocean north of lat. 30°N we find normally in winter a pressure gradient directed northward towards the Gulf of Alaska and the Aleutian Island Region, which is the winter seat of the north Pacific low pressure area, usually referred to as the Aleutian Low. From or around the Aleutian area move most of the winter frontal systems and disturbances which approach the north and central Pacific coast. When these disturbances follow an unusually southerly path, or develop secondary centers which move further south than the usual Pacific depression, or develop in themselves to an unusual intensity, then the Tropical maritime air lying in the northern portion of the Pacific anticyclone is set in motion towards the northeast in response to the intensified pressure gradient toward the north or northwest. Occasionally an extension of the anticyclone northeastward towards the California coast occurs, causing the establishment of a pressure gradient which brings a general flow of Tp air northward along the entire length of the middle Pacific coast even beyond Seattle. Thus in advance of the southern portion of the cold front or the occluded front which marks the southward advance of fresh Pp air behind the depression approaching from the northwest, we may find a broad open warm sector of Tp air moving northeastward or even northward along the coast with rather high wind velocities. Frequently this warm sector is occluded before the warm air reaches the coast, especially when the coastal region is covered with cold continental air. Under these conditions California gets its heaviest winter rain, caused by condensation both within the Tp air aloft, and also within the following unstable Pp air as it overruns the cold continental air (warm-front-type occlusion). The occasional winter development of a marked depression inland as far as the Great Basin or the southern Plateau region of Nevada and Utah, accompanied by widespread heavy precipitation, is apparently associated with marked overrunning at high levels by Tm (either Tp or Tg) air from the south. On the other hand, the Tp air frequently sweeps northward along the Pacific coast with little or no resistance by colder air masses, consequently without appreciable overrunning at the warm front and with little or no rain before the approach of the cold front. A similar development may take place also along the Gulf or Atlantic coasts, but in general the greater warmth and moisture content of the Tg and Ta air masses make the occurrence of moderate to heavy convective showers in the warm current much more probable.

The properties of the Tp air mass may be anticipated to some extent from the nature of the source region. The ocean surface in this region is rather cool for the latitude, ranging in temperature during the coldest season from about 14°C at latitude 35°N to about 19°C at latitude 25°N. The weather condition is prevailingly anticyclonic, which suggests the probability that more or less subsidence is taking place aloft. Thus we should expect to find that the air masses which come from this region are only moderately warm, or even cool for Tropical maritime air, at least near the ground. We should expect to find these air masses also relatively stable, perhaps with indications of a subsidence inversion which should be marked by an abrupt decrease of the w. w should be moderately high near the ground, but definitely not high for a Tropical maritime air mass. Finally, we should expect to observe an abscence of condensation forms in this air until it has moved some distance northward from its source and has been appreciably cooled from below.

However, the data for San Diego in Table V probably give a fairly good picture of the Tp properties in this region. The temperatures we notice are rather high at all levels, though not markedly so considering the proximity of the maritime Tropical source and the southerly latitude. The surface temperature is about 2C° higher than the water temperature off the coast at San Diego, a fact which indicates a moderate recent movement of the air from a warmer region. We notice also the marked stability of the lowest km of the air mass, as well as the large decrease in w, both facts which indicate either a low subsidence inversion or a low turbulence inversion. The rather surprising frequency of low St or Stcu clouds in the TP air mass at San Diego seems to favor the second possibility. If the air mass was originally characterized by marked stability and a rapid decrease of w with elevation at low levels, then the decrease of w at an inversion produced by turbulence might be as large as the difference observed at San Diego between the value at the ground and that at 1 km, otherwise the decrease would not be so large. Above 1 km we note a stable lapse rate, but one which increases somewhat with elevation. The decrease of moisture noted in the first km continues at higher elevations, but especially markedly at the 3 km level.

TABLE V

TROPICAL PACIFIC AIR—WINTER (FEBRUARY 1930)

ELEVATION	SAN DIEGO				SEATTLE			ELLENDALE (TP aloft)		
	T °C	w g	RH %	θ_E °A	T °C	w g	θ_E °A	T °C	w g	θ_E °A
Surface	18.5	9.1	68	316	13.3	7.4	305	(—3.3	2.8	281)
1	16.5	6.1		315	8.0	5.7	308	(+5.3	2.8	297)
2	11.0	5.3		319	1.7	4.3	308	(+1.3	3.7	305)
3	4.5	3.7		318	—2.3	2.9	309	—4.7	2.8	307

In case the water surface in the source region is rather cool for the latitude, as is true for the source region of the TP air masses, then the air mass does not become even conditionally unstable. Thus we note that at San Diego in the TP air mass there is a slight increase of θ_E with elevation, and the same thing is found to be true for this air mass at Seattle. This fact explains why the TP air flow along the Pacific coast gives quantitative precipitation only at an active front. It is too stable for the coast line to be effective in forcing much ascent of the warm current. And even at an active warm front the amount of precipitation from TP air is considerably less than the amount at an active warm front where the more unstable TG or TA air is overrunning.

At Seattle (TABLE V) the general properties of the TP air mass are found to be similar to those observed at San Diego. The air mass has become cooler throughout, but the difference is not very great in view of the 15° difference in the latitude of the two stations. The surface temperature at Seattle in the TP air is approximately 5C° warmer than the lowest average ocean surface temperature which is found during the winter off the coast from this station. The relative warmth of the TP air indicates the rapidity with which this air mass has moved from the south. At San Diego, where the air mass movement is much more sluggish, the difference was found to be only 2C°. But in spite of the warmth of the TP air at Seattle relative to the ocean surface, a fair lapse rate from the ground up is observed in this air mass. The decrease with elevation both of T and of q is much more uniform than it is at San Diego. In particular has the marked stratification of temperature and moisture, which was noted in the first km of this air mass at San Diego, disappeared at Seattle. This is a change which would scarcely have been expected, considering the increasing relative coolness of the underlying water surface with the progress of the air mass northward. It seems to indicate a large degree of turbulent mixing up to a considerable elevation in the TP current. This marked turbulence probably is a consequence of the high wind velocity which must characterize any winter air current of Tropical origin reaching Seattle, and of the ruggedness of the north Pacific coast. But in its general characteristics the air mass is found to be at Seattle as it was at San Diego, warm, moist, and relatively stable. As at San Diego, we note that θ_E increases slightly with elevation, a fact which indicates the absolute convective stability of the mass. Consequently as at San Diego the air mass is characterized by stratiform clouds, and by the restriction of any considerable amount of precipitation to the frontal zones. The characteristic curves on the Rossby diagram for TP air at these two stations are closely parallel and their approximate coincidence with the θ_E isotherms is striking. A comparison of these curves with the characteristic curves of the TG and TA air masses will call to attention at once the comparative instability of the latter air masses, which is indicated by their more nearly horizontal course on the diagram.

The Tg and Ts Air Masses

As would be expected from the uniform warmth of the waters in the TG source region, marked warmth and high moisture content at lower levels is characteristic of these air masses. This is shown strikingly in Table III by the high values of T, w and θ_E at the surface and the 1 km level in the TG air, values which are much higher than are found in any other American air mass in winter. But at Groesbeck we find at the 2 and 3 km levels a surprisingly large decrease in the moisture content of the air mass. This decrease of moisture is best interpreted as due to the presence of another air mass aloft which is designated as S, or TS (Trop. superior), air.

This designation is used for that warm, dry air mass which is found to be present a great deal of the time over the southern U. S. at intermediate and upper levels and frequently over the central U. S. at somewhat higher levels, at all seasons of the year. It is the warmest air mass observed, averaging a few degrees warmer than TG at intermediate levels but, owing to the steep lapse-rate frequently observed in it at higher levels, above about 5 km, it is likely to average slightly colder than TG. It has a relative humidity less than 40%, which is often observed to fall below 20% or even 10% in the upper strata. Normal values (not computed means) of the temperature and specific humidity of this air mass at the 2 km level, at San Antonio, Texas, and Omaha, Nebraska, for January and July, might be given as follows:

	January		July	
Ts at 2 km	T	q	T	q
San Antonio	10°C	3g	24°C	5g
Omaha	5°C	2g	22°C	5g

At higher levels q frequently falls to half a gram or less, and even in summer to less than a gram.

The hot dry weather which occurs in the western Plains states in summer and in the Southwest even in winter, is always associated with the presence of this air mass at intermediate levels and frequently, especially in summer, even at the ground. It occurs frequently above a shallow stratum of moist Tropical maritime air at the surface, but on summer afternoons this moist stratum is likely to be lost by convective and turbulent mixing into the dry TS strata above.

It is the first air mass designation

to be used at M. I. T. which obviously does not apply to a mass of surface origin. For this reason it was clearly identified and designated as a separate air mass only after rather much experience in the use of aerological data in preparing atmospheric cross sections. Formerly it was treated simply as a dry upper stratum characteristically present with the Tropical maritime masses, and not as a distinct air mass. However, it has been found to be of much wider distribution than the Tropical maritime masses, occurring above polar as well as Tropical maritime masses, and obviously requiring separate treatment as a distinct air mass.

In winter at least, the Ts air mass seems to be of tropical origin, as indicated by its warmth, and with an extended past history during which it has remained aloft, as indicated by its dryness. Consequently the logical source region to attribute this air mass to is the upper portion of the sub-tropical belt of high pressure.* We know that slow sinking of poleward moving air must be occurring in this region. This sinking is especially concentrated in the great anticyclonic cells in the sub-tropical high pressure belt, and according to the scheme of J. Bjerknes, especially on the northern and eastern sides of these anticyclonic cellular units. Hence we should expect to find this warm dry air predominating where we have prevailing west or west-northwest winds at intermediate or upper levels in south temperate latitudes. We should expect to find this current at higher levels further north, owing to the increasing prevalence and depth of the polar air masses at higher altitudes. This condition prevails a great deal of the time in the southern U. S., accounting for the prevalence of the Ts air there. Much of this air probably comes to us from the eastern regions of the Pacific anticyclone, after crossing the Rocky mountains.[8] Probably föhn effects on the eastern slopes of the Rockies help to bring the Ts air down to the surface, and to give it the particularly high temperatures observed in the Southwest. However, the appearance of this type of air will be by no means restricted to this region, but should be noted throughout the southern temperate latitudes of the continent.

Recent studies in isentropic analysis by Mr. J. Namias show quite definitely how in summer many of the so-called Ts currents are apparently Pc air masses that have swung around anticyclonically from the eastern Canadian source region and subsided. Thus the heated continent and the seasonal shift of the Westerlies bring the Ts source region well north in summer. (See art. X.)

In spite of the marked potential instability of the Tg air masses in the south, their thermal stability is such that convective precipitation occurs rarely within the air mass in winter. But from the warm front at which the overrunning warm current is Tg air, the precipitation is often convectively irregular and extremely heavy. With the high relative humidity observed in the lowest km of the Tg current, comparatively little forced ascent of the warm current is necessary before the potential instability present becomes available for thermal convection. In fact it occasionally happens even in winter

*See my "Disc. and Illustr. of Problems. . . . of Cross-Sections," *M. I. T., Papers in Phys. Ocean and Met.*, Vol. 4, No. 2, 1935; and G. Emmons: Atmospheric Structure over So. U. S. Dec. 30-31, 1927, *M. I. T. Met'l Course Prof. Notes*, No. 9, 1935 (rev'd *Bull. Am. Met. Soc.*, Aug.-Sept. 1936, pp. 268-271).

[8]The *Ts* designation has supplanted the *Tc* (tropical continental) designation on the North American weather maps, for there is no really extensive *Tc* source region on this continent. The Wea. Bur. uses "*S*" for Ts.

that sufficient forced ascent of a vigorous Tg current is effected by coast line or mountains (southern Appalachians) so that convective showers are initiated, but this occurrence is restricted for the most part to the summer season when insolational heating of the Tg air mass may produce widely scattered heavy convectional thunderstorms in the eastern United States.

In the south the prevailing weather conditions in the Tg air mass are essentially those of warmth and high humidity. There is likely to be a good deal of cloudiness, especially during the night and early morning. The principal cloud forms are rather low stratus or strato-cumulus, which the sun frequently dissipates during the day. Visibility is likely to be rather poor when low clouds prevail but becomes good with the dissipation of the cloud deck. The cloud base is likely to be at only a few hundred meters elevation, while the upper boundary coincides with the temperature inversion which marks the upper limit of the extremely moist air stratum, usually between 12 and 16 hundred meters elevation. It is seldom that any precipitation is associated with this type of stratus in the south, for the inversion above is usually sufficient to prevent penetrative convection and, in its earlier history near the source region, the lower strata of this air current are not sufficiently cooled to produce the mist and drizzle which are likely to appear as it moves northward. Apart from the possible existence of a turbulent stratum near the ground in cases of strong air movement, flying conditions in the Tg air mass should be definitely smooth in the south. Frequent low stratus and rather poor visibility may prove embarrassing, especially in hilly country.

As the Tg current moves northward it is continuously cooled in the lowermost strata by contact with the increasingly cold surface beneath. Over dry land this cooling takes place rather slowly but over cold water or a snow and ice surface it becomes very rapid. The natural consequence is to produce very rapidly a supersaturated stratum at the ground in the warm moist air current. If the air movement is light, this leads quickly to the formation of dense fog at the ground (the so-called Tropical-air fog[9]) but since the advance of Tg air to high latitudes is normally associated with rather strong air currents, mechanical turbulence usually maintains a thoroughly mixed stratum at the ground, topped by a dense low stratus cloud deck. As the cooling continues, rather dense mist or fine drizzle is likely to fall from this stratus, so that the visibility even with rather strong wind may be reduced almost to that of a dense fog, with a ceiling of not more than one or two hundred meters. This is a condition which makes flying practically impossible, except over a perfectly flat surface. Long before visibility becomes poor at the ground the lowering ceiling in the advancing Tg current makes flying in mountainous country difficult. For example, even in the lowest passes in the Appalachians the ceiling quickly closes in, especially on that side of the ranges against which the wind is blowing.

The data in Table III for Tg air at Boston show clearly the beginning of the cooling of the Tg air mass to the Ntm condition, as just described. As compared with Groesbeck we note the marked stability of the first km of

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[9]See H. C. Willett: Fog and Haze, Their Causes, Distribution, and Forecasting, *Monthly Weather Review*, November, 1928.

the air mass at Boston, where obviously the 1 km level is just above the top of the surface turbulence stratum. But equally striking in the comparison of the data at these two stations is the disappearance of the dry Ts strata between 2 and 4 km at Boston. For some reason the moist Tg air mass averages much deeper at Boston than in the southwest. This seems to be a result of the strong flow of the Tg currents in this region and the horizontal convergence which accompanies active cyclogenesis. The dry Ts air mass is usually found here above the 4 km level.

The weather types which we have discussed above in connection with the principal air masses, are the predominant types which we meet with in the United States, in winter, *in the absence of marked frontal activity.* We have seen how the study of the characteristic air mass properties as indicated by the aerological observations facilitates the understanding and explanation of these weather types and consequently favors especially the forecasting of those elements which are important in aviation. Moreover, recent aerological investigation of frontal activity and of the development of cyclonic disturbances here and abroad show that the air mass properties are also of much importance for the explanation and forecasting of all those phenomena which depend upon interaction at the boundary between air masses of markedly different properties (see article by Haurwitz and notes by Bergeron, elsewhere in this booklet).

V. SUMMER AIR MASS PROPERTIES[1]

The Pc and Npc Air Masses

During the warm season the source properties of the American Pc air masses are very different from those characteristic of the cold season. The snow cover which is always present over this source region during the colder half of the year completely disappears, and the bare ground surface becomes much warmed during the long summer days. Consequently instead of being cooled from beneath as in winter, the Pc air masses are heated from the ground. Even though the air mass may originate over the cold waters of the Arctic, it quickly loses the coldness and stability of its lower strata over the warm land. It is usually impossible to tell from its properties whether the air mass came initially from the ocean or not. This holds for the Pacific as well as for the Arctic Ocean, for in any case by the time the air mass has crossed the western mountain ranges or moved southward from the Arctic Ocean to the field of observation it has acquired about the same characteristics. These characteristics are typically as follows (*Cf.* Figs. 16-18 of Art. VIII-IX by Namias) :—

(1) A fairly low moisture content (low compared with that of summertime air masses of more southerly origin). The values of w at the ground usually are near 5 or 6 grams, and decrease steadily with elevation. Relative humidity is also low, especially during the warmer time of day, so that conditions are favorable for the steady supply of moisture by evaporation from the warm ground and its transport upwards by convection. It is probably in this way that the observed moisture distribution has been established. With increasing age and displacement southward of the air mass the w values increase steadily at all levels from the ground up.

(2) A moderately low tempera-

[1]Excerpts and abstracts by R. G. Stone from "American Air Mass Properties," 1933.

TABLE VI. THE Pc AIR MASSES—SUMMER

Elevation Above Sea Level (km)	Ellendale T °C	Ellendale w g	Ellendale θ_E °A	Royal Center T °C	Royal Center w g	Royal Center θ_E °A	Pensacola (NPc) T °C	Pensacola (NPc) w g	Pensacola (NPc) θ_E °A
Surface	19.0	6.3	312	17.0	8.3	314	22.8	13.4	332
1	16.2	5.6	313	12.8	5.8	311	19.5	9.8	330
2	9.8	3 9	312	5.8	4.5	310	12.2	7.2	325
3	4.0	3.1	314	2.0	2.6	311	7.0	5.0	324
4	—2.5	2.9	318	— 1.8	1.4	314			
5				— 7.2	1.0	318			
6				—15.5	1.3	320			

ture. The Pc air mass in summer is characteristically cool, compared with masses of southern origin, but hardly to be distinguished in temperature from the mP (Pp) masses of the Pacific. The dryness of the Pc mass favors a large daily temperature fluctuation in its lower strata near the ground, for this condition is conducive to radiational cooling by night and insolational heating by day. In general, as it proceeds southward, the air mass tends to become increasingly unstable. However, since nearly all the upper air data available are for the early morning hours, this instability is not apparent in the data shown in Table VI. In the middle west a diurnal temperature range at the ground of approximately 15°C is typical for this cool dry air mass, a range which is usually sufficient to establish a dry adiabatic or superadiabatic lapse-rate up to 2 or 2½ km by early afternoon.

(3) A lack of condensation forms. Due to the dryness of the summer Pc air mass, it is typically cloudless. In spite of the marked daytime instability of the mass, the condensation level is so high that only a few scattered high Cu clouds at most are likely to be observed. In the later portion of its life history, however, the conditions favoring rapid evaporation of moisture from the ground to the Pc air mass may so increase its moisture that eventually even thundershowers may develop locally. This is, however, unusual, occurring only in cases of marked stagnation of the air mass movement.—*Excerpts.*

[The Ellendale data show the properties of the air that has come SE over western Canada; there is a moderate lapse-rate with distinct stability near the ground which later in the day (data are early a.m. ascents) will probably approximate the dry adiabatic up to 2 or 2½ km and often exceeding it in the first km. Up to 3 km w decreases steadily and θ_E is constant (but increases above that). At Royal Center the general trend of the vertical distribution of these elements is similar; however, lower levels are distinctly colder and moister here in spite of the fact that the latitude is 6° farther south. At higher levels it is warmer and dryer at Royal Center than at Ellendale. These differences result from the fact that the Pc air at Royal Center has passed over the cold Hudson Bay waters while the Ellendale Pc air has come over the warm northwestern Canadian interior. It is characteristic of summer Pc outflows, that instead of showing properties of a discrete mass of cold air rapidly breaking out from some cold air reservoir in the north as they do in winter, they usually appear more in the nature of almost stationary deep northerly currents (more NW-ly in winter) that may remain in the same longitudinal zone for several days. Thus Royal Center in summer does not, as in winter, get Pc air that has passed first over Ellendale, but rather directly out of the north, with the resulting differences mentioned above.

At Pensacola, Fla., the maximum modification (to Npc) of the original Pc properties may be seen; note the high w and some decrease in θ_E with elevation, which indicate conditional instability, especially in the afternoon, so near to actual instability that local thundershowers may develop. The surface temperature is not far from the lower limit for Tm air in summer (340°).—*R. G. S.*]

The Pp Air Masses

Generally speaking, we expect to find in summer all maritime air masses relatively stable, and continental masses relatively unstable, compared with their winter vertical structure, because of the seasonal reversal of the normal temperature differences between land and water surfaces. We have seen that for the Pc air masses this difference is very pronounced, that the condition of extreme stability of the winter continental Polar air is changed in summer to a condition of moderate instability, especially marked during the daytime. But a glance at the summer properties of the Pp air mass at Seattle at upper levels (Table VII) shows the presence of a surprisingly good lapse-rate, especially through the first km. Since all the Seattle ascents were made during the early morning hours, this surface instability cannot be explained as the result of insolational heating during the short interval that the air has been moving inland from the sea. It must be explained rather as the result of turbulent mixing effected in an air mass initially moderately stable by its passage over the mountainous promontories or along the devious water route to the head of the fjord where Seattle is located. The constancy of w, and the increase of the relative humidity from an average surface value of 62% to an average of 91% at 1 km is further evidence of the correctness of this assumption. Stcu clouds are nearly always present with a base elevation between 8 and 14 hundred meters under these conditions. But the Cu-nb clouds and showers characteristic of the Pp air mass in winter are definitely absent. The reason is obvious when we note a lapse-rate between the 1 and 3 km levels of only four-tenths of the dry adiabatic rate, whereas in winter between these levels we found a lapse-rate of more than seven-tenths of the dry adiabatic rate. The large decrease of w to be noted in the Pp air mass in summer between the 1 and 2 km levels is noteworthy as indicating definitely that between these levels must lie the upper limit of the turbulence layer and doubtless therefore of the Stcu cloud layer also.

TABLE VII. SEATTLE Pp—SUMMER

Elevation Above Sea Level (km)	T °C	w g	RH %	θ_E °A
Surface	16.5	7.1	62	308
1	8.5	6.3	91	308
2	4.5	3.9		307
3	0.5	2.3		308.5
3½	—2.5	1.7		310

In general, however, in spite of the fact that fresh Pp air at Seattle is characterized by greater stability between 1 and 3 km than is summer Pc air at the inland stations, nevertheless, no condition approaching isothermalcy is found at any level in the Pp air masses, but rather quite an appreciable lapse-rate at all levels. Furthermore, we find that from 1 km upwards the Pp air mass is markedly colder than the Pc mass at any of the inland stations. The temperature found at 3½ km in fresh Pc air at Seattle is the same as that observed at 4 km in Pc air at Ellendale, where this air mass is colder aloft than at any of the other inland stations. Furthermore, the values of w at Seattle are slightly less than those in the Pc air masses at Ellendale and Royal Center. Rather noteworthy is the marked constancy of equivalent-potential temperature in the Pp air at Seattle. This constancy of θ_E indicates the absence of potential instability at any level of the Pp air mass, but on the other hand it definitely does not represent the stable structure of an air mass cooled from below.

The coldness and dryness of the summer Pp air masses at Seattle indicate almost conclusively the correctness of the assumption, based on the study of the winter properties of this air mass type, that the source of the air mass is the same as that of the Pc mass, or at least that it is as truly Polar or Arctic in character. In winter we found that although the lower strata of the Pp air mass were greatly warmed and moistened by the warm ocean, the upper strata were quite as dry and only a little warmer than the same strata of the Pc air mass at Ellendale. In fact, recent observations extending to higher levels indicate that above 4 km the Pp air masses are frequently, and even in winter, colder than the Pc air. In summer we find the Pp air mass cooler than the Pc mass at all levels, and above the first km just as dry. The relative coolness of the maritime Polar air mass in summer depends primarily upon the coolness of the ocean surface relative to the land surface. Not only are the lower strata of the maritime air mass less heated by their contact with the earth's surface than are the same strata of the Pc mass, but there is also less terrestrial radiation from the cool water surface to the upper layers of the maritime air mass. The absence of the marked heating by contact which occurs in winter at the warm ocean surface is reflected also in the low elevation to which the dampness of the summer Pp air mass extends, i.e., the small vertical extent of the penetration of mechanical or convective turbulence. Nevertheless, the lapse-rate in the summer Pp air masses at Seattle is steep enough to indicate that the air flow has been from colder to warmer surfaces. Probably the heating is rather gradual during the progress of the air mass from the cold Polar seas southward. This heating probably becomes effective at upper levels only by means of direct radiation from the surface beneath without mechanical or convective turbulence having played any role above an elevation of about 1 km, which we found to be a normal depth of the turbulence layer at Seattle.

But there is besides the relative coolness of the water surface in summer another fact which should not be overlooked in the explanation of the coolness of the Pp air masses at this season. This fact has to do with the change in the normal atmospheric pressure distribution along the north Pacific coast from winter to summer. In summer the pressure is relatively low over the continent, and relatively high over the ocean, especially over the northern area, so that in summer the middle Pacific anticyclone extends far northward into the region normally occupied in winter by the so-called Aleutian Low, which tends to be displaced inland from its winter position with greatly diminished intensity. Consequently in summer there is normally prevalent a well-marked pressure gradient directed from ocean to continent along the entire Pacific coast from California northward, a condition which favors a steady transport of Polar maritime air southeastward along the entire Pacific coast. This condition is so persistent in summer that warm maritime air from the south seldom if ever reaches the north Pacific coast at the surface directly, as it frequently does in winter. Occasionally there occurs a temporary cessation of the maritime Polar outflow on the north Pacific coast during the passage further north of a disturbance following a more southerly course than is usual in summer. Furthermore, the ocean surface temperature along the California coast is so low in summer, because of the upwelling of cold water, that Tropical maritime air masses must pass a great distance northward from their source region to reach the latitude of Seattle.

It seems very probable that the change at San Diego from the Seattle values of the Pp air mass properties would be in the direction of a marked increase in stability, especially at the surface. Probably even in the prevailing air flow from the northwest we would find a pronounced low turbulence inversion (the wind velocities are usually too great to permit of the formation of a surface temperature inversion) with dense St or Stcu clouds, which may at times approximate surface fog in their low elevation in the upper portion of the turbulence layer. The extreme local coldness of the ocean surface along the California coast is sufficient to cool the lower strata of even the coolest Pp air masses from the northwest.

It is quite impossible in summer to distinguish air of Polar Pacific origin from that of Polar continental origin after the former has reached the aerological stations in the interior of the U. S. In summer most outflows of Polar air over North America first become evident in a strengthening of the normal pressure gradient along the Pacific coast, and consequently in an intensification of the Pp air flow. This northerly current gradually works inland, with an accompanying general rise of pressure in the coastal region, and a gradual displacement eastward of the zone of strongest north or northwest winds. Consequently the Polar air current becomes increasingly continental in its composition as the Polar source region from which the outflow takes place is displaced continually inland, or eastward. Under these conditions it becomes almost impossible to determine a boundary between the air current of maritime and that of continental origin. In winter this distinction is easier to make, because of the very much greater coldness and dryness of the continental air. But in summer when the initial differences between the Pp and Pc air mass properties are so slight, by the time that the Pp mass has crossed the mountains, usually rather slowly, and come probably into radiation equilibrium with the continental surface beneath, all differences between these air masses are so completely obliterated that it is neither possible nor of any advantage to distinguish between them. Consequently the designation Pc can be used indifferently, in summer, for Polar continental air masses or Polar air masses of Pacific origin after they have reached the interior of the U. S. —*Excerpts.*

The PA Air Masses

The Polar Atlantic air masses are more important in the late spring and early summer than at any other time of the year. At this season the ocean surface in their source region, from Cape Cod northeastward to Newfoundland, is at its maximum of abnormal coldness for the latitude and at its coldest in comparison with the continental region to the west. This relative coldness is a consequence of the slowness of the cold continental ocean current from the north in warming up in the spring, and possibly in part to the drift of ice into this region with the Labrador current. It follows that this ocean region in late spring and early summer becomes a real cold air source. Whenever the normal eastward movement of air over this region ceases, as it frequently does in the spring and summer, the tendency is towards the immediate development of a stationary anticyclone thermally maintained by the cooling of the stagnant air mass present in the region. Such developments frequently manifest surprising persistence over the cold water, and usually lead eventually to an overrunning of the north Atlantic coastal region by the cold PA air of the anticyclonic circulation. Occasionally it happens that a general southward movement of the cold air follows down the entire Atlantic coast as far south as northern Florida, bringing with it a decided drop in temperature. The difference in temperature between this cold maritime air and the hot continental air may amount to as much as 20° or 25°C. Usually, however, the PA air masses in summer do not make their influence felt south of Cape Hatteras, and the greater part of the time only on the north Atlantic and especially the New England coast.

The term PA, in summer as in winter, is applied to those air masses which were originally Pc, but which have remained long enough over the cold waters of the north Atlantic to have become appreciably modified. We have seen that in winter very little time is required to effect such a modification because of the marked initial coldness of the air mass. In late spring and early summer, however, the water surface is colder than the surface strata of the Pc air, so that the modification takes place slowly. On the other hand, the general stagnation of the air movement over this north Atlantic area is frequently so persistent at this time of year that the air mass may have days in which to reach a condition of equilibrium with respect to the surface beneath. As the stationary maritime anticyclone develops under these conditions, the cold air usually reaches the coastal stations at first as not much more than a sea breeze, but on the following days the cold air mass usually invades the whole coastal area east of the Appalachian Mountains, and occasionally advances far to the south. The properties of the cold air mass are shown best by the New England coast stations, although in the case of a southward displacement of the air mass the characteristic coldness is retained to a surprising degree. The surface air temperature of the mass, as indicated by an outlying station like Nantucket, is probably very close to that of the cold ocean surface from which it is moving. This temperature is likely to be about 5°C at the beginning of May, about 10°C at the beginning of June and 12° to 15°C later in the summer. There is almost no daily period in these temperatures. In spite of the fact that these temperatures indicate a cooling of the air mass from the temperature which it originally possessed over the continent as Pc air, the Pa air mass is found to have just as in winter a rather unstable structure up to about 1 km. Since this lapse-rate cannot be explained in this case as caused by heating from below, mechanical turbulence remains as the only obvious explanation of the instability of the lower km of the Pa air mass. Stratiform cloud forms indicate a marked inversion at the top of the turbulence layer. It is very probable, in view of the prevailingly stagnant anticyclonic condition associated with this air mass, that the inversion is intensified by continual subsidence of the upper strata of the mass. The wind velocity is usually strong enough in the cool maritime anticyclone to justify the turbulence explanation of the unstable ground layer of the Pa air mass, and the long exposure of the mass over the cold water could account for the loss of much heat from this stratum by turbulent transfer downward to the cold surface. Two April ascents at Boston in this type of air mass showed a lapse-rate nearly nine-tenths of the dry adiabatic rate up to 1 km, where there was a thin Stcu cloud layer, with a temperature inversion immediately above of 5°C. We find typically in the Pa air mass in summer some St or Stcu or Frst clouds at the top of the instability layer, though they are usually much thinner than in the same mass in winter, seldom covering the entire sky, and frequently appearing as only a few scattered Frcu or completely disappearing. Precipitation never falls from these clouds in summer. This follows from the initial dryness of the Pc air mass, and the very small amount of evaporation which takes place from the cold water in summer. There is very little difference between the specific humidity of the Pc and that of the Pa air mass at this season. The cooling of the air mass at the cold water surface, an effect which we assume to be carried upward by turbulence, produces usually a thin saturated stratum at the top of the turbulence layer (as indicated by the cloud formations mentioned above), but seldom more than about 70% relative humidity at the surface. Hence the real Pa air mass does not become foggy over the water, but is on the contrary usually characterized by excellent horizontal visibility, apart from the occasional thin cloud layer mentioned above. There is, however, occasionally visible over the ocean on a clear afternoon in this air mass a very noticeable whitish haze, even when the relative humidity is appreciably below saturation, and the condition is far from being one of real fog. This may be due to the presence of tiny water droplets on salt nuclei.

It is obvious that the general effect of the underlying cold water surface in the source region of the Pa mass in summer is to effect a cooling of the air mass from beneath making it essentially stable, though this is apparently counteracted close to the ground by turbulence. Once the Pa air comes over the warmer land it rapidly becomes warmed from below and hence more unstable.

It should be mentioned at this point that warm, moist, continental air in summer, and even more Tropical maritime air masses from the Gulf Stream, which move into the

PA source region, are very quickly cooled over the cold water to such an extent that dense fog is immediately formed. Partly because of their greater initial warmth and moisture, and perhaps partly also because of less wind and mechanical turbulence, the dense fog appears almost immediately at the surface and grows deeper with prolonged cooling. It is this condition which gives to this region its reputation for spring and summer fogginess. These air masses of Tropical origin, cooled in the PA source region, are designated on the M. I. T. maps as NTM. The symbol NPA may appear on the summer weather maps for PA air which has moved far south over warmer water, or been brought to a considerable distance inland.—*Excerpts.*

The Tropical Pacific Air Masses

During the warm season the Tropical Pacific air masses play a negligible role on the Pacific coast. It was remarked in the discussion of the PP air masses in summer that the pressure distribution at this season along the entire middle and north Pacific coast of North America is such that there is found a persistent onshore gradient, and a correspondingly persistent movement of air at least to a considerable elevation from the north and northwest. This pressure gradient may temporarily disappear to such an extent that the air movement for a time is almost stagnant, the winds becoming light variable. At such times the aerological ascents at Seattle indicate the presence of higher temperatures at all levels, and usually of somewhat higher specific humidity at the ground, than occur in the PP current when it is well developed. But at upper levels one usually finds a dryness indicative either of subsidence or of a light continental outflow of air with some föhn effect from the mountains inland. During the summer of 1930 there was not found a single instance either of marked southerly air movement or of a moisture distribution indicative of TP air at Seattle at the ground. Probably this particular summer was quite typical of the average summer in this respect, so that it may be concluded that real TP air does not appear on the Pacific coast of the U. S. in summer, at least at low levels. This can certainly be said of Seattle. At San Diego the situation is much more uncertain, as the condition there is normally one of comparative stagnation. Probably it would be somewhat warmer and somewhat moister in summer than in winter, corresponding to the higher temperature of the ocean surface in the source region. On the California coast this air mass, if it appears at all, should be definitely cool at low levels and foggy, because of the extreme local coldness of the water along the coast.

It seems quite probable, however, that in the Plateau and Rocky Mountain regions the TM air is of real importance in summer. Temperatures in the Plateau region are rather high at this season, and the pressure normally rather low. Under these conditions rather steady southerly winds frequently are observed in this region for rather prolonged periods, with the result that the moisture thus brought inland establishes a specific humidity high enough so that considerable cloudiness and widespread thundershowers develop throughout the low pressure trough. If the air thus brought into the Plateau region had its source in the comparatively cool and dry PP current prevailing along the coast, an extremely low relative humidity would prevail in the air in this heated inland Plateau which is normally too dry to furnish much moisture by evaporation. We have found that Pp air in summer is characterized by a specific humidity on the order of 7 g at the ground, decreasing to only slightly over 2 g at 3 km elevation. Warm air which reaches Ellendale from the northern Plateau low pressure region is observed to have a specific humidity of from 10 to 12 g at 1 km elevation, a value which definitely suggests that the air is of Tm origin. Recent studies, notably by Reed, of the upper winds in summer in the Southwest, indicate that this warm moist air moves frequently at the upper levels from the Tg source region via Mexico. It is quite possible that it may come also from the Ts source region occasionally, but apparently it is predominantly Tg air.

The Tropical Gulf, Tropical Atlantic and Ts Air Masses

For the Tg and Ta air masses the normal summer conditions are much more favorable than they are for the Tp air masses. The combination of the tendency toward the development of low pressure over the interior of North America and the tendency toward the development of a well-marked center of high pressure over the western Atlantic Ocean (Bermuda High) results much of the time in summer in a pressure distribution which brings mT air northward over most of the eastern U. S. and even into Canada. Consequently the mT air masses in the eastern and central U. S. are present a much greater part of the time and extend over much wider areas in summer than they do in winter. They are responsible for the oppressive heat with high humidity which more than anything else characterizes our summer weather in the eastern and central U. S. The map of July 26, 1930, 8 a.m. represents almost in ideal type form the general condition which leads to the widespread prevalence of mT air in the U. S. in summer. We notice a weak trough of low pressure over the northern Plains states, and a broad extension of the Bermuda High westward over the southeastern U. S. The resulting general air movement consists of a light flow from the West Indies and Caribbean Sea northwestward into the Gulf of Mexico, thence northward over the southeastern and south central U. S., and thence northeastward into the Lake Region and towards New England. In other words, we observe a slow steady flow of Equatorial air which originated in the Trade-wind zone over the entire eastern and central U. S. As this condition is usually very stationary, the stream lines on this map as indicated by the isobars may be taken as typical trajectories of the mT air masses in summer, the air masses to which the following discussion applies.

The general difference between the normal winter and the normal summer condition as regards the distribution and prevalence of the mT air masses may be expressed in another way by saying that the zone of maximum frontal activity between the mT and cP air masses, or the sub-Polar front, is displaced in summer from its normal winter position somewhere over the northern Gulf of Mexico, northward into the U. S. almost to the region of the Great Lakes.

In general the properties of the mT (Tg and Ta) air masses in summer as they leave their source region are similar to their properties in winter in the same region. The ocean surface temperatures during the warmest season average over the entire Gulf and Caribbean Sea region close to 28° or 29°C. In the Caribbean Sea this is only 3°C warmer than the temperature during the coldest season, a difference which increases probably to nearly 10°C on the immediate Gulf coast. Thus we should expect the mT air mass in summer to leave the source region somewhat warmer and somewhat moister than it does in winter, but quite similar to its winter condition in its general vertical structure. We should also expect to find that as the mT air mass passes inland from the source region the general tendency of the effect of the continent will be towards a raising of the air mass temperature, instead of towards a lowering, as it is in winter.

Pensacola, Fla., doubtless gives the best indication of the source properties of TG air in summer. We note a surface temperature in the TG air which is almost identical with the water surface temperature of the source region. Above the surface we find a moderate lapse-rate, about 0.6 of the dry adiabatic rate, which indicates a condition of thermal stability in the air mass. Afternoon ascents would doubtless indicate a steeper lapse-rate near the ground. The relative humidity is surprisingly high, in view of the warmth of the air mass, so that the specific humidity exceeds slightly the large value of 20 g. Probably this value is as great as would be found as an average in any maritime location in Equatorial regions. Not only is w extremely high at the ground at Pensacola, but the values found up to at least the 3 km level indicate very high relative humidities at the prevailing temperatures. At the same time the values of w observed at Pensacola at all levels in the TG air are higher than those found at any other station. In spite of the high moisture content found at the 3 km level, the excessive amount of water vapor present in this air mass at low levels gives rise to a condition of marked potential instability. Up to the 3 km level a decrease of 18° in θ_E is noted, while the high value of w at this level would indicate that the decrease in θ_E should continue for at least 2 km further. This condition implies that all convective or mechanical turbulence up to at least 5 km elevation must effect an upward transport of latent heat, and the high values of the relative humidity indicate that very little vertical displacement is necessary in order to initiate active convection which should extend well beyond this level. This same marked potential instability which we observe in the TG air masses coming from the Gulf of Mexico and the Caribbean Sea in summer, and which we shall find presently to be characteristic of all maritime Equatorial air masses, is the source of the great amounts of energy consumed in the genesis and maintenance of the Tropical hurricane. When we consider the added effect of insolational heating of the TG air mass near the ground during the daytime as it moves inland from the Gulf coast, it is obvious why Cu clouds which develop during the afternoon into Cunb and heavy local thundershowers are so characteristic of this air mass.—*Excerpts.*

[In Texas the invasion of summer TG is typically shallow, with dry TS air aloft above 1 or 2 km, even less sometimes, which makes thunderstorms in TG much less likely than farther east; the decrease in θ_E aloft indicates much potential instability but in the absence of much moisture aloft it is not realized except from marked vertical displacements with passage of occasional pronounced warm or cold fronts.*—*R. G. S.*]

Modification of the Tg and Ta Air to Ntm

As the TG air mass moves northward from the Gulf of Mexico in summer we should not expect as a rule any very great changes in its properties. During the late spring and the summer the North American Continent is insolationally warmed even at rather high latitudes. Only in the region of the Great Lakes and off the north Atlantic coast is there any possibility of a marked cooling of the TG air mass from the surface beneath. Apart from these water areas the ground is normally warmer, especially during the daytime, than is the water surface in the source region of the air mass, so that the general tendency must be towards a warming of the air mass from the warmer surface beneath as the TG current moves northward from the Gulf of Mexico.

At greater distances from the source region we would expect to find a small decrease in the moisture content of the TG air mass as the result of the precipitation of some of the excessive moisture in widely scattered thundershowers. Especially in the west and at upper levels we should expect to find the influence of the gradual infiltration and overrunning of dry TS air from the upper levels of the sub-tropical anticyclones.* But in general, a condition of oppressive warmth and moisture with afternoon thunderstorms should characterize MT air over the entire eastern and central U. S. The temperatures at all levels tend to run a little higher than at the Gulf coast stations. This difference is, perhaps rather surprisingly, a maximum at the 1 and 2 km levels. This is to be explained by the early morning hour of the majority of the ascents, a fact which is indicated by

TABLE VIII. THE TROPICAL GULF AIR MASSES—SUMMER (1930)

Elevation Above Sea Level (km)	Pensacola				Broken Arrow				Ellendale				Royal Center			
	T °C	w g	RH %	θ_E °A	T °C	w g	RH %	θ_E °A	T °C	w g	RH %	θ_E °A	T °C	w g	RH %	θ_E °A
Surface	28.5	20.7	85	359	29.5	15.4	58	348	28.5	16.5	66	351	29.0	15.9	61	349
1	23.2	15.6		350	26.5	12.3		345	27.0	13.3		346	25.2	13.9		348
2	17.0	12.5		345	20.2	9.9		342	21.8	8.7		339	18.5	11.5		344
3	10.8	9.5		341	12.8	8.2		339	12.8	5.7		332	10.8	8.6		339
4					7.0	5.4		336								
5					1.5	3.5		336								

*See discussion of Ts air in summer on pp. 94-95 of this booklet; and in the article on isentropic analysis by Mr. Namias, as well as in the note Showalter at the end of this section.—*Ed.*

the stability of the lowest km of the air mass at the inland stations, the consequence of nocturnal radiational cooling. The heating of the upper strata of the air mass takes place during the afternoon thermal convection, but during the night the temperatures near the ground fall.—*Excerpts.*

[The highest temperatures aloft occur at Ellendale, farthest from the Gulf, and indicate the heating effect of the continent to perhaps 5 km or more. Stations inland all show lower w at all levels and, except in the southeast, where there is direct invasion of TA air, lower w at the ground, than at Pensacola. North from the Texas coast the Ts and TG layers become increasingly mixed by diurnal convection, which lowers w and θ_E at the surface and increases them aloft. Eastward of the western plains the influence of Ts air decreases at the surface as continental heating becomes less marked. As in winter, it is difficult to distinguish TA and TG air masses.

The pure or slightly modified TM summer air in the southeastern United States is so conditionally unstable and humid to well above 5 km that thunderstorm convection is frequent; this becomes, as we have seen, steadily less marked inland to the north and especially westward, where the dry and stable Tropical Superior (Ts and modified TG) air masses with their clear skies, good visibility, and large diurnal temperature range stand in decided contrast to the humid, hazy, showery and cumulus-laden TM air.—*R. G. S.*]

In summer the modification of the typical TG or TA air mass to the NTM condition by cooling from below occurs only in rather restricted regions, for over the continent proper such cooling does not take place. It occurs only over a cold water surface. On the Canadian side of the Great Lakes this effect is sometimes noticeable, especially in the early summer when the Lakes are still cold. At this season the mT air from the south may arrive on the northern shores of the Great Lakes definitely cooled and even foggy or with low St clouds. But this cooling effect is most important over the cold water off the north Atlantic coast. In only a day or two of slow movement of the warm moist mT air mass over this cold ocean surface, the lower strata are cooled almost to the water surface temperature, and the cooling effect is carried upward, presumably by mechanical turbulence, with surprising swiftness. The resulting condition is one of stable stratification with the rapid formation of dense fog of considerable depth. This is the cause of the high frequency of summer fog on the north Atlantic coast. It is principally in this region that the NTM designation is used on the synoptic charts in summer.—*Excerpt.*

Major Frontal and Air Mass Zones of the Earth

Bergeron describes nine latitudinal *Weather Zones* of each hemisphere as a function of the large scale distribution of Air Masses and Fronts*) (see *Met. Zts.*, v. 47, p. 249, figs, 1, 6, and *Ymer*, hf. 2-3, 1937, p. 218, fig. 10):—

1. The *Stratus zone of the Stable Polar Air* (or Arctic Air*) in its source region.
2. The *Shower zone of the moist-labile Polar Air* (or Arctic Air*) in somewhat lower latitudes.
3. The *Dry zone of diverging Polar Air* (or Arctic Air*) underneath the Polar Front surface (or Arctic Front surface*).
4. The *area of continuous precipitation from Nimbostratus at the Polar Front* (or Arctic Front*).
5. The *Stratus zone* (with or without drizzle) *in the Stable Tropical Air at the Polar Front* (or in the Polar Air at the Arctic Front*).
6. The *first Cumulus zone of the Tropical Air* in middle latitudes.
7. The *Dry zone of diverging Tropical Air* in the high pressure belt of the horse latitudes.
8. The *second Cumulus zone of the Tropical Air:* the trade-wind zone.
9. The *Shower zone of the moist-labile Tropical Air* in the equatorial region (the Doldrums).

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*When there is also an Arctic Front polewards of the Polar Front, five more Weather Zones are arranged along the former front as along the Polar front.

Further Studies of American Air-Mass Properties

(Excerpts from *Mon. Wea. Rev.*, July, 1939, pp. 204-217)

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WINTER SEASON

Continental Arctic Air: As shown by Willett and Wexler cA air is probably formerly maritime polar air (mPK) which is cooled by surface radiation forming cAW. A study of this air mass by Wexler has shown that there is a very sharp inversion near the surface, and above, a lapse rate approaching the isothermal. Since the effect of surface cooling rarely extends above 3 km above sea level, the uncooled air above would still be mP and very few observations of cA air are available above that level. Occasionally cA air is mechanically lifted to 4 km at Cheyenne and in such a case it will be cooled adiabatically and a temperature very low for that height will result. Unstable Arctic air, cAK, sometimes occurs in the Hudson Bay and the Great Lakes regions but insufficient data are available to include cAK air in this study.

The change in designation from cAW to cPW has been more or less arbitrary but usually one or two days elapse before cAW air is considered cPW. A more definite criterion for the change in notation is the formation of a new Arctic front. Eventually the original polar air may become tropical air, so cPW merely marks the transitional stage.

The striking thing in the movement of cAW - becoming - cPW air into the southern United States is that the steepening of the lapse rate which would be expected from the addition of heat from below does not occur in the mean, except in the lowest few hundred meters. Apparently subsidence proceeds so rapidly at all levels above the shallow turbulent-convective layer that the great stability characteristic of cAW air near its source is still preserved and cAW - becoming - cPK is rare, except behind deepening cyclones. As the polar air feeds into low latitudes it spreads out to occupy several times its original area and thus the compensating subsidence shown by the various mean cross-sections and tables of average properties [see original article] is accounted for.

There seems to be considerable moisture added in the lower levels by surface evaporation but because of the extreme stability of cAW and cPW air it does not seem likely that the effects of surface addition of moisture extend to any appreciable elevation. Consider for example, the mean value of 2.2 g/kg for specific humidity with a potential temperature of 288° A at 2 km at Oklahoma City in cPW air. To establish an adiabatic lapse rate to carry such a quantity of moisture up to 2 km by vertical convection, a potential temperature of 288° A is required at the surface. Since a surface potential temperature of 288° A is found only in air of very nearly tropical properties it is evident that the observed amount of moisture could not have been carried up to that elevation by simple vertical convection. The addition of moisture at higher levels in modified polar air is probably best explained by the principle of horizontal mixing along isentropic surfaces as discussed by Rossby (see following art. on Isentropic Analysis).

Maritime Arctic Air, MA:—When an outbreak of polar air moves over only a very small part of the Pacific Ocean before reaching the United States it is usually designated as MAK. If its trajectory has been far to the south, it usually is sufficiently modified to be called mP.

Strictly speaking, for air masses entering the United States, the notation mAk should be reserved for Arctic air masses which move directly southward along the North Pacific coast and have only a short trajectory over the ocean. The notations mPk and mPw are adequate to differentiate between maritime polar outbreaks having longer trajectories (see figure 1). The unusual instability of mAk air, some flights indicating that vertical convection has obtained to at least 6 km, has two important effects on its interaction with other air masses in the central and eastern part of the United States. First, since often it is colder aloft than the surrounding air masses, sinking from these levels, or subsidence, occurs in mA and mP air. Second, and inversely, this same instability in mA air is apt to cause an increasing tendency for vertical divergence and increasing instability in air masses moving into a region occupied or recently occupied by mA air. In other words, the mA gains in stability while the surrounding masses lose.

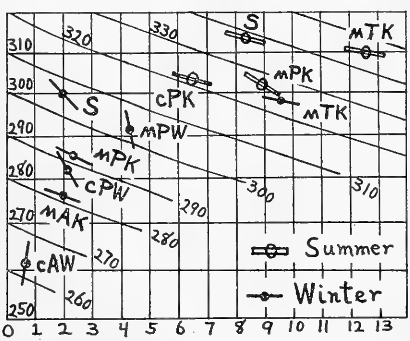


Fig. 1. Average θ_E and Slopes of Characteristic Curves for the levels between 1000 and 1500 meters for American air masses, based on U. S. soundings of 1935-6.

During the winter months mA air near the surface is usually warmer than the continent and shortly after passing the coast line it should be labelled as a warm-type air mass according to the Bergeron classification. Over snow covered areas mAk air very rapidly assumes continental characteristics; thus we find the modified form of mA air sometimes colder near the surface than the original.

Maritime Tropical Air, mT (Tm): Since mT air is formed out of air which was originally polar, it shows a tendency for vertical stability with some subsiding action in the higher levels. It will be noted that although the partial-potential temperature of the dry air increases fairly rapidly with elevation, the equivalent-potential temperature usually decreases with elevation. This seems to indicate that at higher levels this air mass is relatively dry. Since mT air usually moves inland with anticyclonic motion it may be assumed that the relative dryness aloft in mT air can be explained by horizontal divergence with subsidence in the upper levels of the subtropical anticyclonic cell. There is evidence that some of the water vapor at higher levels must have been carried upward by vertical convection over scattered areas in the Gulf and Caribbean, and was diffused to the surroundings. The evident upward slope of the isentropic surfaces to the northward indicates that the moisture is carried aloft not only by vertical convection but also by isentropic mixing. The upward transport of moist air appears to occur near the edges of the subtropical anticyclones, while the subsiding dry tongues originate nearer to the centers.

Superior or Subsiding Air, S (Ts): The notation Ts (Tropical superior) was originally applied to air supposed to have been derived from the upper subsiding portions of the subtropical anticyclonic cells. However, of recent date the designation S has been applied to all *warm* air masses that show relative humidities below 40%, which is taken as an indication of subsidence and horizontal divergence. The study indicates that most of the dryness results from the subsidence of high-level air from a polar source. Isentropic analyses have shown definitely that in winter a number of dry tongues move out from polar regions.

Modified moist Polar air, MPW (NPM): The mean values for these air masses show a tendency for decreasing stability and increasing moisture content. Although there is evidence that considerable heat and moisture are added by surface effects, there must be an addition of moisture aloft by isentropic mixing.

SUMMER SEASON

Continental polar air, cA–becoming - cP (Pc - becoming - NPC): The rate of increase of moisture aloft seems slightly in excess of that possible through vertical convection and it is necessary to believe that horizontal mixing plays an important role.

Maritime polar air, MA - becoming - MP (PP - becoming - NPP): The effects of subsidence and outflow from the upper portions of MP air seem to have about the same prominence as the effects of horizontal mixing so the mean values for MP air do not give positive evidence of the increase of moisture aloft by isentropic mixing.

The notation cP or NP should be used in cases of doubtful history of the polar current, or in cases of overlapping layers of maritime and continental air having been thoroughly mixed by vertical convection during the day or in some cases by mechanical turbulence. Since there is so little difference in the properties of MP and cP after 1 day's history over the continent during the summer months, it would be wise to label all polar air masses cP in the summer time, after they have moved east of Spokane and south of the Canadian border.

Modified moist polar air, MP (NPM): The rapidly increasing moisture seems to be a combination of the effects of surface evaporation over water surfaces or areas with abundant plant life, coupled with the effects of horizontal mixing. It seems possible therefore for a polar air mass with a purely continental history to attain in summer quantities of heat and moisture comparable to those obtained by an air mass moving over the Caribbean Sea.

Maritime tropical air, MT (TM): It appears that in general the potential temperatures are higher in MT air than in continental air masses and that therefore the air would move upward along the isentropic surfaces as it came over the land. Some of the cases studied indicate that at times a given isentropic surface may slope downward from the Gulf to the continent. This means that portions of the MT column may at times move downward in approaching the continent. It will be noted that mean potential-temperature surfaces are approximately horizontal between Miami and Pensacola.

Taking the vertical distance in meters between the 303°A and the 311°A surfaces as an inverse measure of stability, one finds the MT air to be more stable in the mean than MPK air but less stable than cPW air. This agrees with the statement made above to the effect that MP air was likely to cause increasing instability in air masses moving into its territory. The author is of the opinion that the greatest probability of rain occurs when MT air replaces MA or MP, or when MA or MP replace MT air. The latter sequence is more conducive to precipitation during the colder seasons.

Superior or subsiding air, (S): The assumption that the dryness of this air mass is due to subsidence is not necessarily always correct in summer, because rapid surface heating of relatitvely dry air at that season may produce a deep column of air whose moisture content is far from the saturation values. When such an air mass is cooled during the night by surface radiation some slow sinking may occur in the layers near the surface and those layers which are not cooled by radiation will show low relative humidities by the time of the airplane observation the next day. Since S air is recognized only on the basis of relative humidities less than 40%, the source of this air can therefore be traced to subsiding polar or tropical air or to rapid daytime heating of either of those air masses.

The range in specific humidity and equivalent potential temperature is considerable at all elevations for the dry type of air called S, so it can be assumed from this evidence also that the source of S air may be either polar or tropical or a stratification of air from both sources with a tendency for horizontal mixing along the isentropic during the night and vertical convection during the day resulting in a dissipation of the concentration of moisture. The orographic effect also plays an important role in the development of S air east of the Rocky Mountains.

Air Mass Classification

Our study indicates that it is possible to make some reasonable standardization of air mass classification for synoptic purposes but that any classification falls far short of definitely identifying the thermodynamic properties of the different air masses. In other words the mean values of each of the various properties (see figs. 12 to 15 of original article) show definite groupings for the different air masses, and the individual values show definite limits for the properties of *fresh* tropical and *fresh* polar outbreaks. However, the modifying influences affecting air masses, especially in the summer months, result in a very wide range of equivalent-potential temperatures as indicated on the frequency distribution charts (figs. 5 and 6 of original article). It can be seen from a study of these charts that the classification is relative for any one day, and no definite limits of equivalent-potential temperature have been in use. This seems regrettable for statistical purposes but in practical synoptic work it can hardly be avoided.

It is usually the practice to label the air masses differently on either side of a front, and since polar air is sometimes modified very rapidly, it happens that modified polar air behind a cold front on one weather map may have a higher equivalent-potential temperature than a modified tropical current behind a warm front on a map a week later. Since all tropical air is polar air modified by surface effects, any classification of air masses must be only a compromise as to number of types. The author is of the opinion that no purpose is served by increasing the number of types.

The Air Mass Cycle

It is possible to identify from the mean seasonal properties for the different air masses, two distinct cycles of transition from polar to tropical air, one a moist cycle with rapid addition of moisture, the other a dry cycle with a marked subsidence and slow isentropic mixing in the early stages.

For the winter season the moist transitional stage (figure 8, original article) from mP as represented at Dayton, to Tm at Pensacola and St. Thomas, Virgin Islands, shows continual subsidence and increasing moisture. The persistent stability throughout the transitional process, the difference in potential temperature between 500 m and 5,000 m remaining practically constant from the mA to the Tm stage, indicates that the principal addition of moisture must occur by means of mixing along isentropic surfaces, with some probability of convection in the final mT stage.

The dry winter cycle (fig. 9 of original article) shows rapid subsidence with slight increase of moisture by isentropic mixing from the mA to the S stage. When this air mass moves to lower latitudes the subsidence decreases, rapid surface heating develops and the S air begins to mix both vertically and horizontally with mT air. The vertical mixing is probably confined to the lowest layers affected by daytime convection and most of the increase in moisture aloft appears to be due to isentropic mixing.

The identification of the moist and dry cycles is more difficult in the summer season because of the greater tendency for vertical convection during the day, with convection occurring under saturated conditions which cannot be analyzed by charts using potential temperature surfaces. However, the mean values, indicating unsaturated conditions, suggest for the moist cycle conditions similar to those observed in the winter season, namely, subsidence and surface heating with rapid increase of moisture by isentropic mixing. Appreciable quantities of heat and moisture may be added to mT air over the continental United States. Thus the highest value of equivalent potential temperature, 366° A, observed during this study was found at 1,000 m at Dayton on August 22, 1936!

The dry cycle in summer represents rapid modification of the polar air masses with subsidence aloft and slow addition of moisture by isentropic mixing over the continent, then continued heating accompanied by vertical convection and convergence over the Gulf and Caribbean, followed by a slow spreading out and subsidence as the air assumes an anticyclonic trajectory on its return from the lower Caribbean to El Paso. From El Paso to the Mississippi Valley there is apparently a continued addition of heat and moisture and the mT air again becomes convectively unstable over the Mississippi and Ohio Valleys.

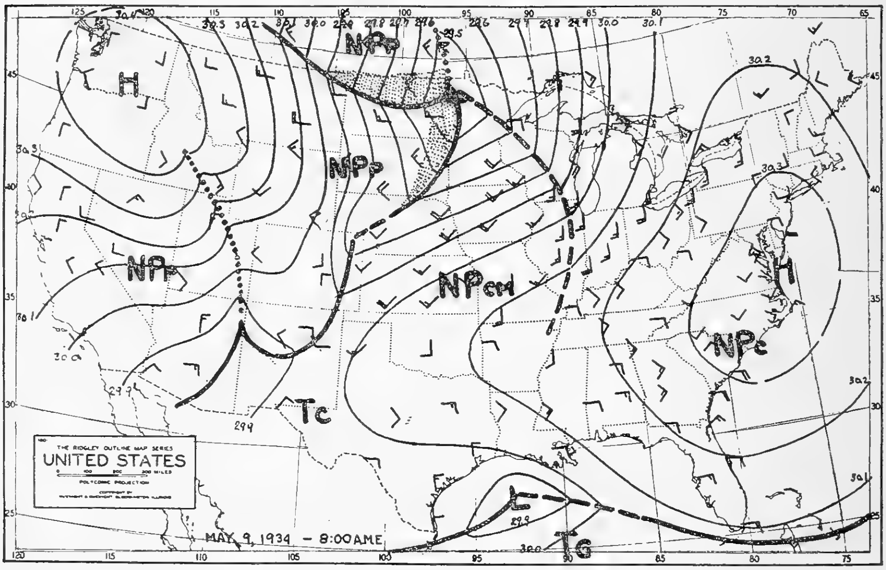
In view of recent discoveries of the meteorological significance of isentropic charts it is further recommended that more attention be given to the slope of potential temperature surfaces in situations free from condensation. Allowance should be made for the effects of horizontal mixing along isentropic surfaces and unless the isentropic surfaces in one air mass actually intersect the ground or at least show a sudden increase in slope, the synoptic analyst should label the air masses differently with caution.—(*Excerpts.*)

Illustrations

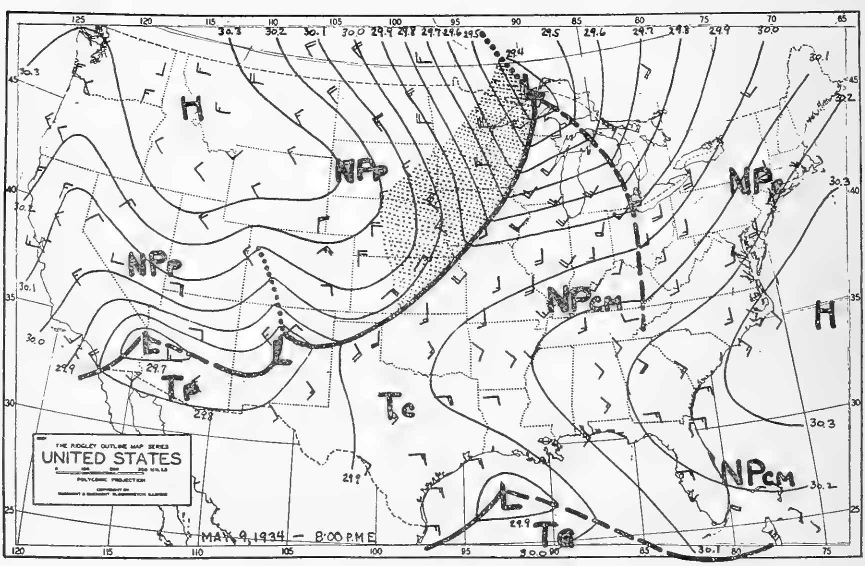
The following collection of weather maps and cross sections selected from articles published in the Bulletin of the American Meteorological Society is reprinted to offer examples of frontal and air-mass-analyzed weather situations, though the technique of presentation used is not altogether conventional. But they will give those who have no access to other publications nor to MS analyzed maps of a meteorological service or institute, some idea of interpretations of atmospheric structure and movements that can be or are made by experienced map analysts.—*Ed.*

A Dust Storm

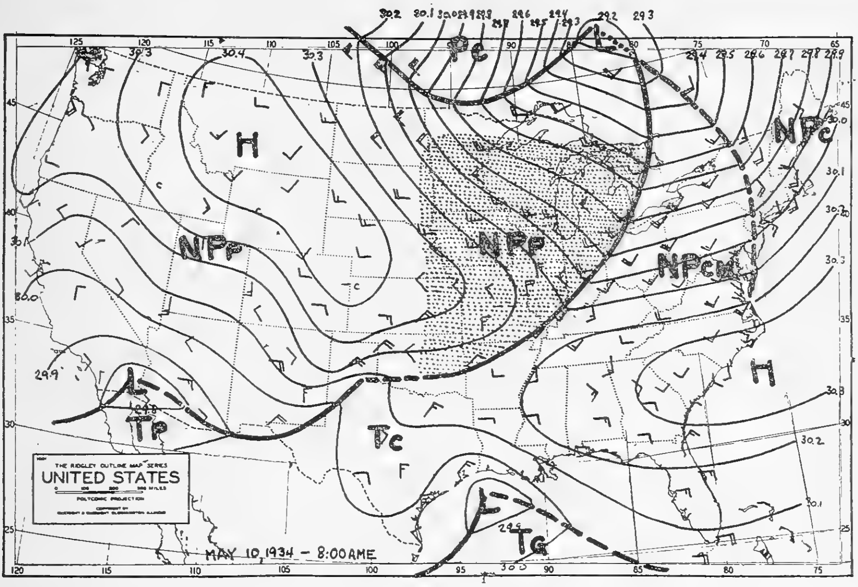
May 9 to 11, 1934, a famous duststorm raced from Montana and North Dakota to the Atlantic Seaboard and out to sea, one of the few such to reach the ocean. The maps for this situation, below, were analyzed and discussed by G. R. Parkinson (BULL., May, 1936, pp. 127-35.) The dust storm generally rages in NPP air, which is unstable (by day at least) and windy in the winter and spring. Note how in this case the dust (stippled areas) was raised in the foreward part of NPP air masses where the pressure gradient was steep and hence the wind fairly high and the lapse-rate unstable, and in passing over regions of barren soil due to drought, plowing and winter; once raised to considerable heights in the air mass, the dust stayed in suspension long enough to be carried still higher and a long ways East.



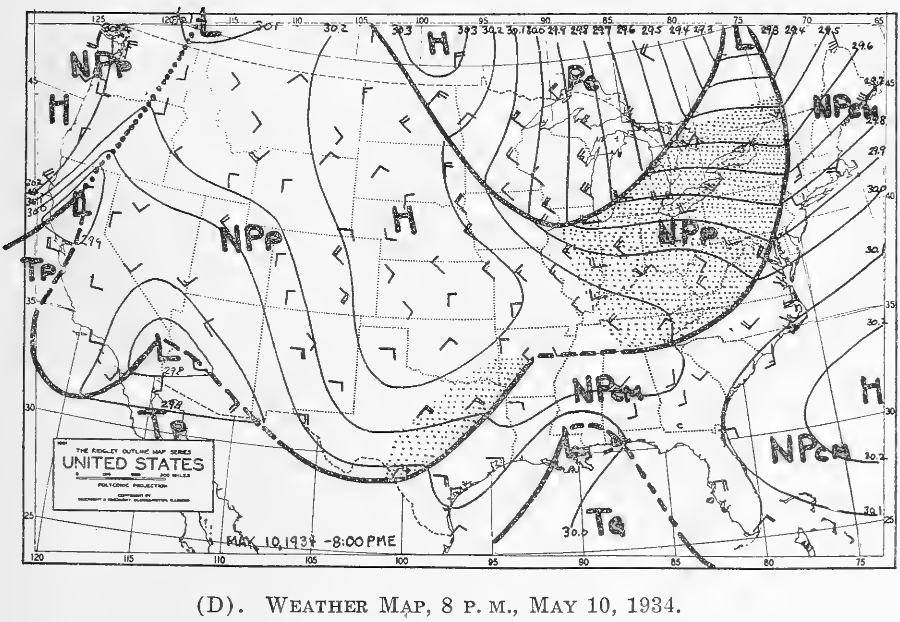
(A). WEATHER MAP, 8 A. M., MAY 9, 1934.



(B). WEATHER MAP, 8 P. M., MAY 9, 1934.

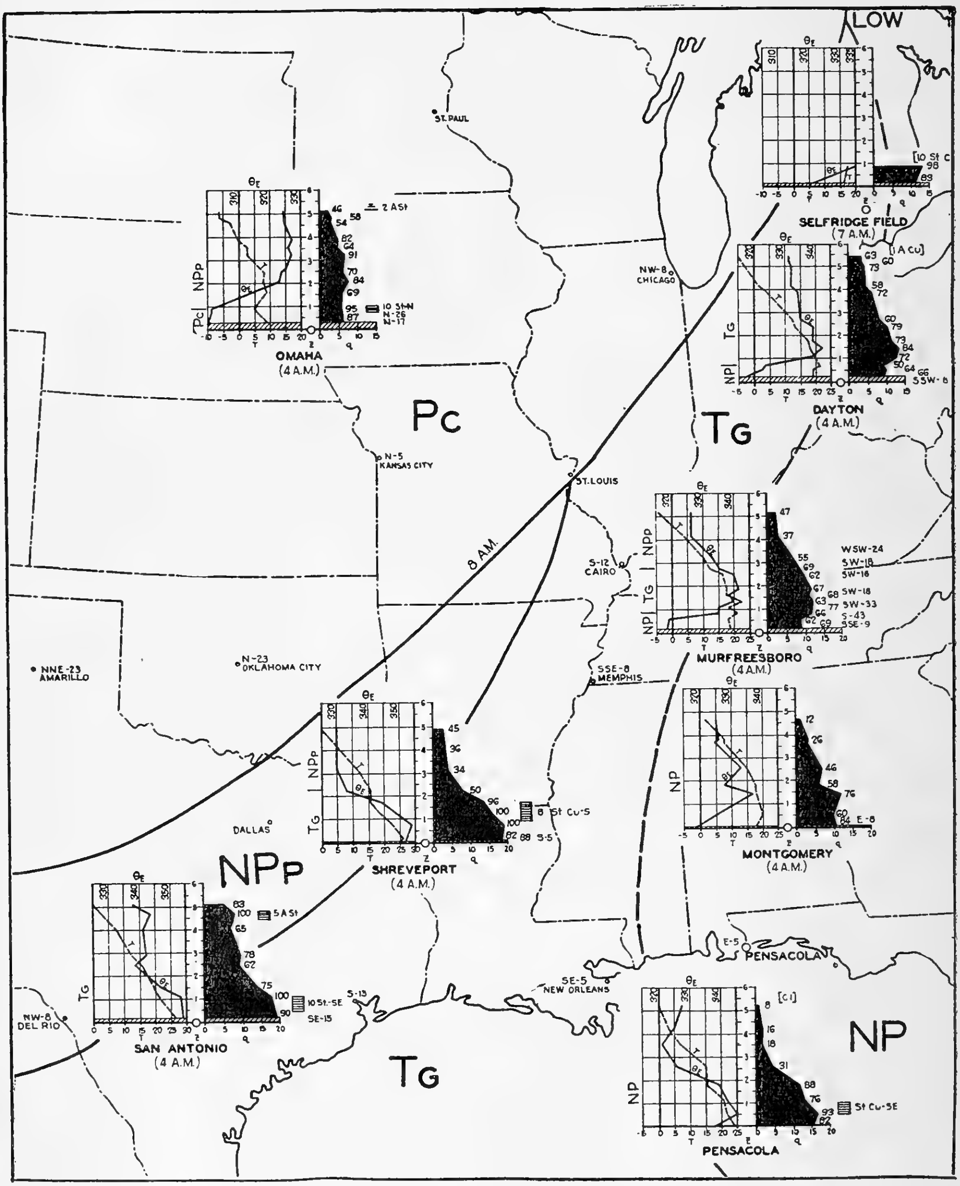


(C). Weather Map, 8 a.m., May 10, 1934.



(D). Weather Map, 8 p.m., May 10, 1934.

Flood Rains



LOCATION OF FRONTS AT THE SURFACE AT 7:00 A. M., C.S.T., SEPT. 27, 1936, WITH DIAGRAMS OF THE CHARACTERISTIC PROPERTIES OF THE AIR MASSES ACCORDING TO THE AIRPLANE SOUNDINGS OF EARLY THE 27TH, (E.S.T.). (Solid lines are cold fronts, dashed lines warm fronts, dotted lines occluded fronts. The graphs show temperature in °C, θ_E, and specific humidity, plotted against height in km above sea level, with air mass types, cloud layers, winds and

(cont., next page)

relative humidities written in figures at the sides; winds are for the 6 a.m. C.S.T. pilot-balloon runs. The Montgomery sounding is inserted for reference.)

This unconventional diagram by E. J. Minser (Jan. BULL., 1938, p. 34) gives an interesting display of upper-air data, most of the surface data being omitted to avoid confusion. The situation was one in which very heavy rains fell in eastern Texas, Oklahoma and Mo. as the PC and NPC cold fronts slowly squeezed (occluded) the moist TG sector ahead of them.

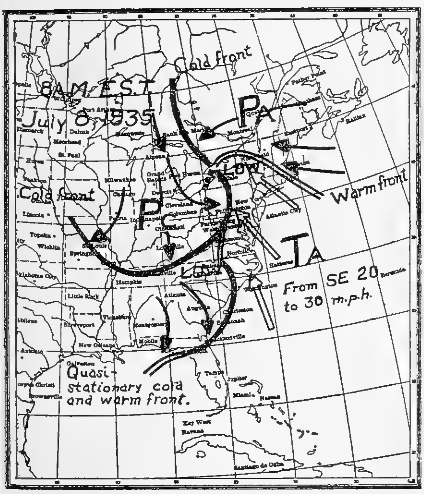


FIG. 1. MAP OF 8:00 AM, EST, JULY 8, 1935.

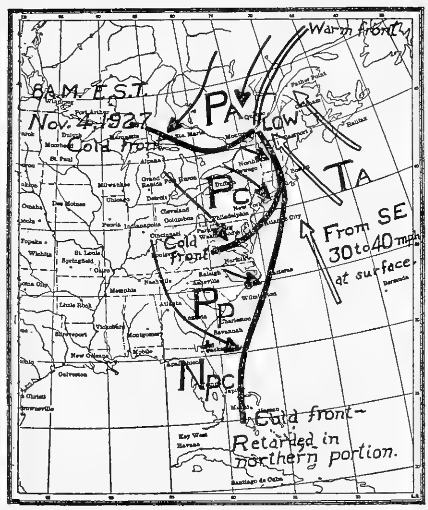


FIG. 2. WEATHER MAP OF 8:00 AM, EST, NOVEMBER 4, 1927.

Two great flood producing rainstorms situations are analyzed here in outline by H. R. Byers. Note how in each case the set up was similar: a more or less stalled N-S cold front with moist TA air streaming over it rapidly and giving heavy rain under the front and against the mountains for many hours or days. (BULL., March, 1937, pp. 128-36). The March, 1936, and May 31, 1889 flood situations were similar too.

Fronts and Aircraft Icing

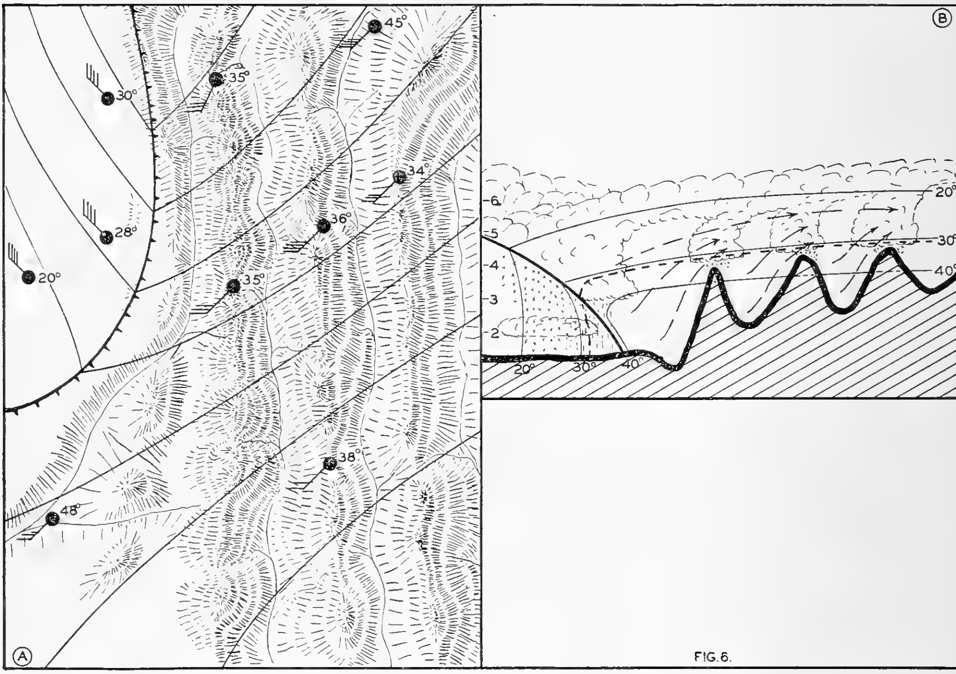


FIG. 6.

Cold Fronts.—This (Fig. 6) and the following cross-sections (Figs. 7-9) through fronts are based on the synoptic experience of analysts working for a large air line (T.W.A.), and are taken from a paper on situations for icing of aircraft by E. J. Minser (Bull., March, 1938, pp. 118-121), with his kind permission. These meteorologists obtain radioed reports from the pilots as they fly along, and thus acquire an unusually detailed knowledge of actual cloud and upper-air structures. Fig. 6 is semi-schematic but vividly illustrates what happens just ahead of an advancing cold front (Pc into Npc) in a mountainous region: clouds with snow squalls (this is winter) over each range. Note the freezing-line (dashed), the frontal precipitation (first rain, then snow), and the surface map at left. (Temp. in °F, heights in thousands of feet, in this and following figures.)

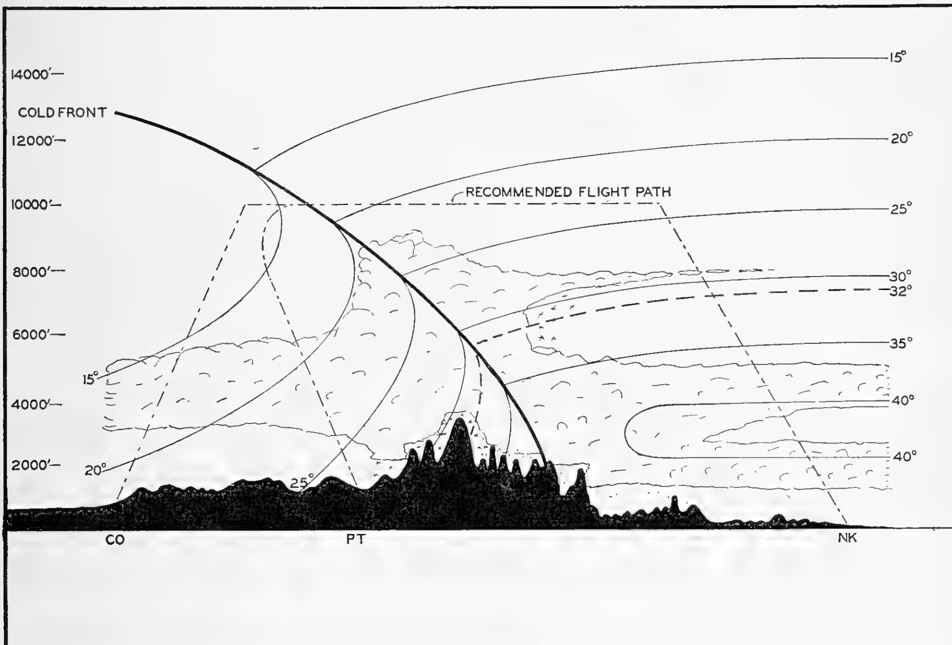


FIG 7

FIG. 7 is generalized cross-section from Columbus, O., (CO) to Newark (NK), showing a typical cold front with aircraft icing conditions in its clouds. The dashed-dot lines show how a pilot must fly to avoid the worst ice forming dangers. Snow squalls in the mountains are probably from both the colder and warmer air. (The windward sides of the mountains are more apt to be fogged in with clouds under such conditions than clear as shown here.)

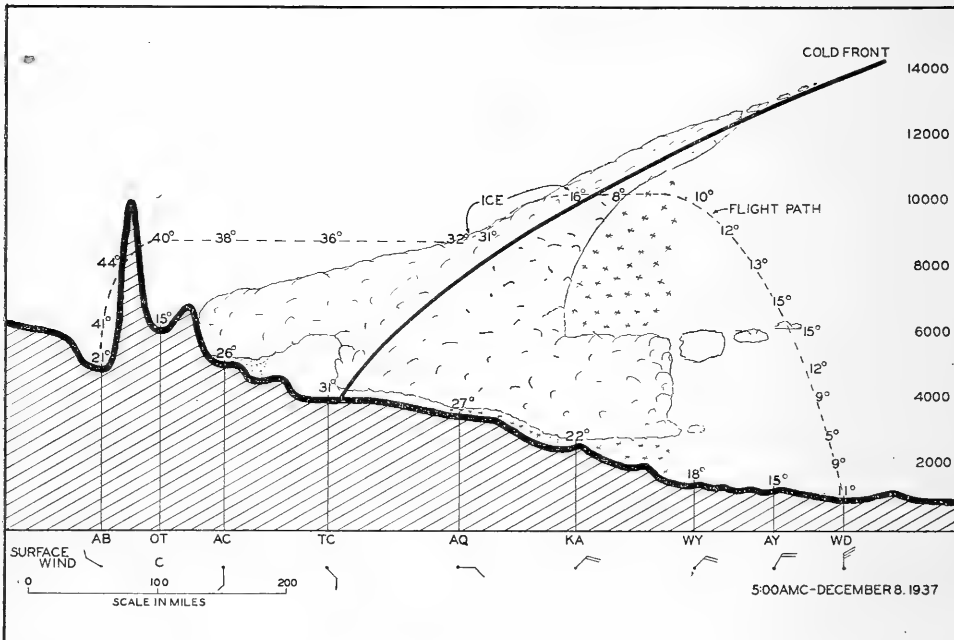


FIG. 8.

FIG. 8 is from actual reports of pilots flying between Albuquerque and Kansas City in the morning of Dec. 8, 1937. A Pc cold front is advancing westward and lifting the Npc air over it and over the mountains. Snow squalls (indicated by crosses).

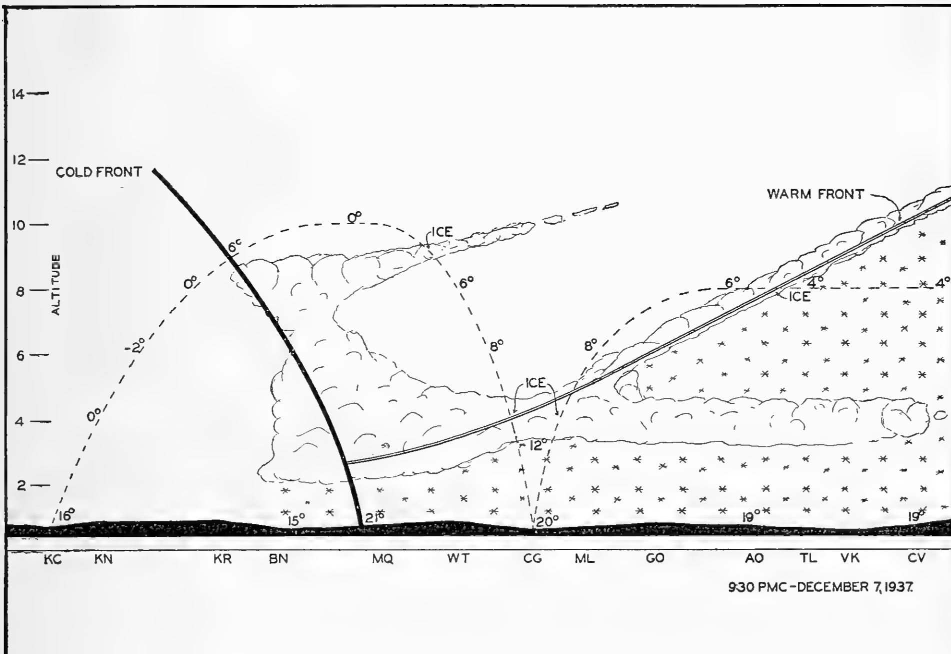
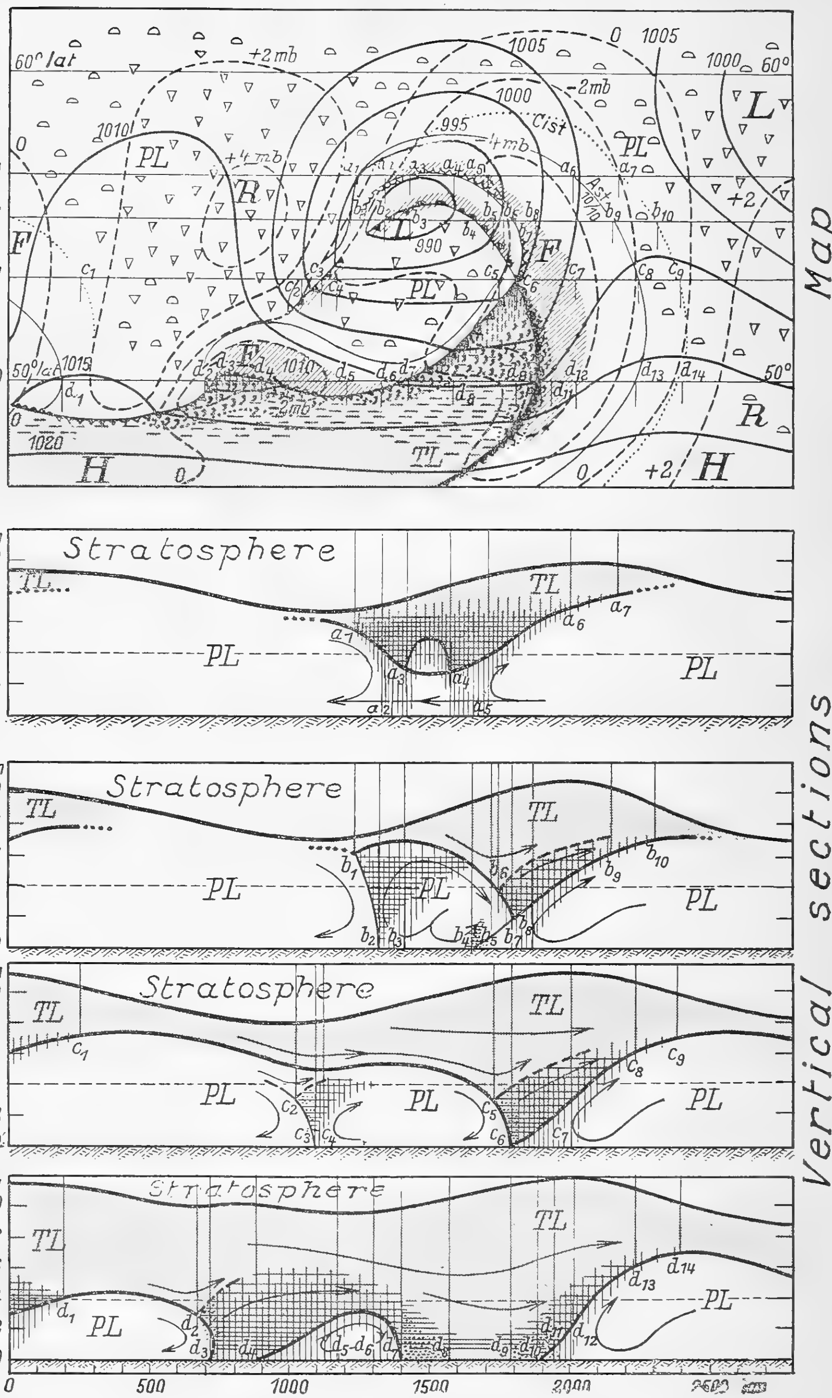


FIG. 9.

FIG. 9 is the actual section from Kansas City through Chicago to Cleveland based on a pilot's flight at night. There is a cold-front-type occlusion with the warmest air-mass entirely aloft here, causing snow to fall over a wide area.

Bergeron's Model of the Warm-Front-Type Occlusion

(Characteristic of Western Europe, Western North and South America, temperate zones)



Legends for Model of the Warm-Front-Type Occlusion

On the map:—

- Cumulus
- Showers
- "Upslide" Rain
- Limit of Ast 10/10
- Warm-sector Rain
- Drizzle
- Stratus or Fog
- Warm Front
- Cold Front
- 1000 Isobar
- +2 Isallobar
- *In the high pressure wedges:* Limit of Cist
  Along the fronts: Limit of drizzle

In the vertical sections:—

- Surface of Polar Front or of Tropopause
- Internal Frontal Surface
- Occluded Frontal Surface
- Stream-lines referred to a system moving with the occlusion
- Cloud-mass consisting of ice-crystals
- Cloud-mass consisting of water-droplets
- Lower limit of ice nuclei
- Ordinary precipitation
- Drizzle
- Warm air

PL = Polar Air. TL = Tropical Air

Characteristic Points of the Occlusion Model Indicated in Fig. 4.

Vertical Section A:

a_1 = Rear limit of Ast, attached to the recurving Occlusion
a_2 = Rear limit of Nbst, attached to the recurving Occlusion
a_3 = Upper retrograde Cold Front, becoming Warm Front
a_4 = Upper progressive Cold Front
a_5 = Fore limit of Nbst attached to the original Occlusion
a_6 = Fore limit of Ast attached to the original Occlusion
a_7 = Fore limit of Cist attached to the original Occlusion

Vertical Section B:

b_1 = Rear limit of Ast attached to the recurved Warm Front, now Cold Front
b_2 = Recurved Warm Front at the ground, now Cold Front
b_3 = Fore limit of Nbst, attached to the recurved Warm Front, now Cold Front
b_4 = Occluded progressive Warm Front at the ground
b_5 = Fore limit of the new Upslide Cloud System, formed on the occluded Warm Front
b_6 = Rear limit of the old Upslide Cloud System, formed on the original Warm Front and modified by the Upper Cold Front
b_7 = Upper progressive Cold Front
b_8 = Fore limit of Nbst formed on the original progressive Warm Front Surface
b_9 = Fore limit of Ast formed on the original progressive Warm Front Surface
b_{10} = Fore limit of Cist formed on the original progressive Warm Front Surface

Vertical Section C:

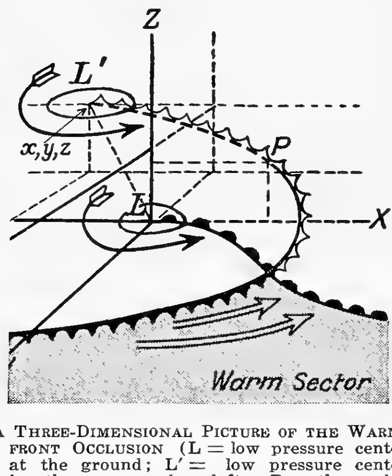
c_1 = Fore limit of Cist of the next disturbance
c_2 = Rear limit of Acu *opacus* attached to the secondary Cold Front
c_3 = Induced secondary Cold Front
c_4 = Fore limit of Nbst formed by the secondary Cold Front

(*Continued on next page*)

c_5 = Rear limit of Acu *opacus* attached to the primary Cold Front
c_6 = Occlusion Point
c_7 = Fore limit of Nbst formed on the primary Warm Front Surface
c_8 = Fore limit of Ast formed on the primary Warm Front Surface
c_9 = Fore limit of Cist formed on the primary Warm Front Surface

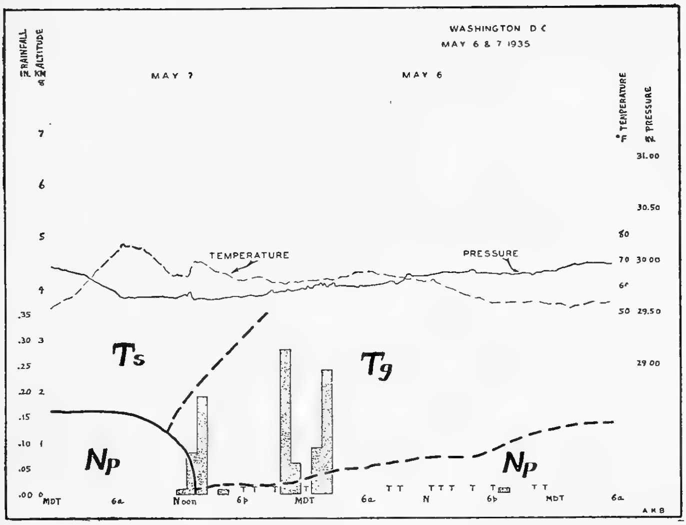
Vertical Section D:

d_1 = Fore limit of Ast of the next disturbance
d_2 = Rear limit of Acu *opacus* attached to the wave disturbance of the Cold Front
d_3 = Cold Front of the wave disturbance on the primary Cold Front
d_4 = Warm Front of the wave disturbance on the primary Cold Front
d_5 = Fore limit of Nbst on the Warm Front Surface of the wave disturbance
d_6 = Rear limit of Nbst formed on the primary Cold Front Surface
d_7 = Primary Cold Front
d_8 = Limit of drizzle before the primary Cold Front
d_9 = Limit of drizzle behind the primary Warm Front
d_{10} = Primary Warm Front
d_{11} = Limit of drizzle (= rear limit of Nbst) before the primary Warm Front
d_{12} = Fore limit of Nbst formed on the primary Warm Front Surface
d_{13} = Fore limit of Ast formed on the primary Warm Front Surface
d_{14} = Fore limit of Cist formed on the primary Warm Front Surface

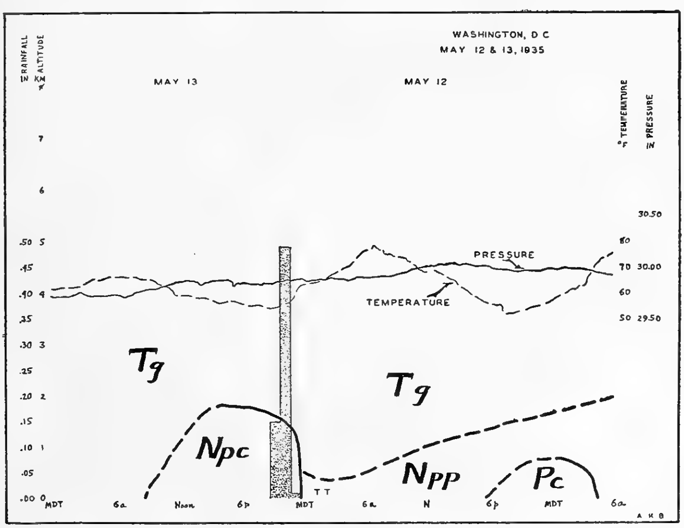


A THREE-DIMENSIONAL PICTURE OF THE WARM-FRONT OCCLUSION (L = low pressure center at the ground; L′ = low pressure center in the warm air aloft; P = the point where the upper cold front pierces the x, z-plane).

Spring Showers



AIR MASS AND WEATHER SEQUENCE PASSING OVER WASHINGTON, D. C., MAY 6-7, 1935.—A time-line cross-section in which the earliest hour is at the right. This unusual diagram was made up from a study of the surface observations and the synoptic cross-sections at the Weather Bureau in Washington for the purpose of summarizing the climate of the whole month in terms of air masses and fronts. However, it incidentally also shows an interesting typical sequence in the upper air. Note the showery rain (stippled bars) from the TG being occluded and lifted between NP and older NP; and the location of TS, which is characteristic—the rain stops abruptly when it arrives. The pressure and temperature traces at the ground are typical. There was a thundershower at the cold front. The TG sector is of course cloudy, the TS/NP zone nearly clear. (From: Botts, Sept., 1937, BULL. p. 290-297.)



Another of Bott's sections, showing pronounced cold-front rain from a cold-front-type occlusion.

Winter Cyclone with Dust Storm

(From: Parkinson, BULLETIN, May, 1936, p. 130-1)

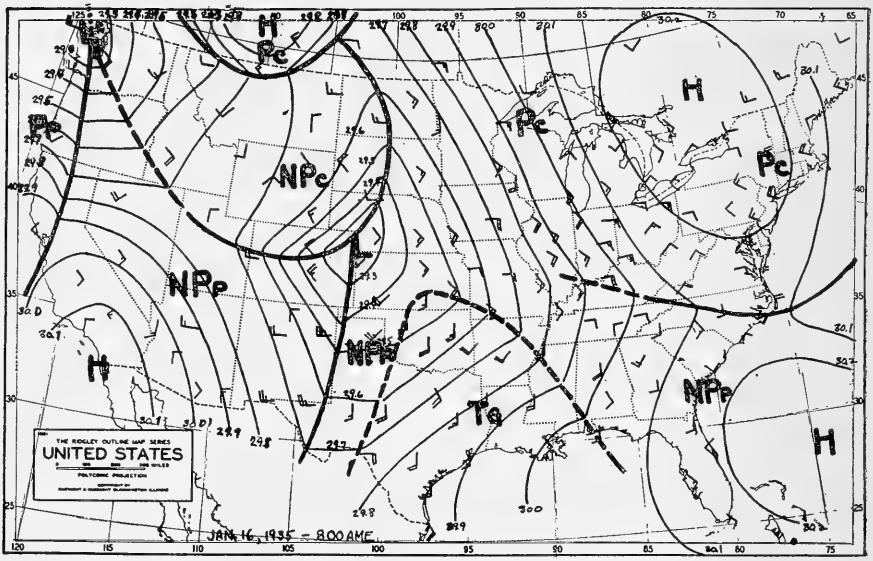


FIG. 1. WEATHER MAP, 8 A.M (75TH MERID. TIME) JANUARY 16, 1935. (Solid lines=cold fronts, dashed lines=warm fronts, dotted lines=occluded fronts, stippling=duststorm areas.)

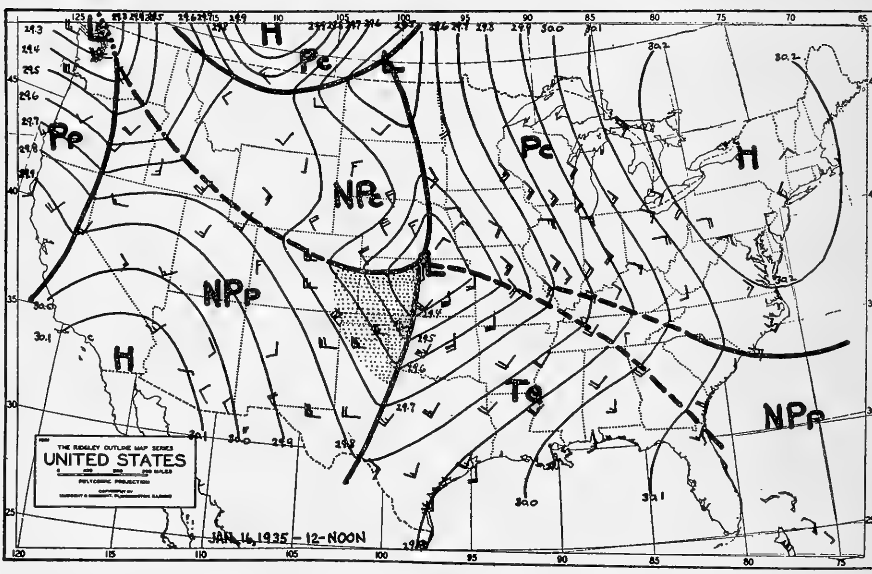


FIG. 2. WEATHER MAP, 12 NOON, JANUARY 16, 1935.

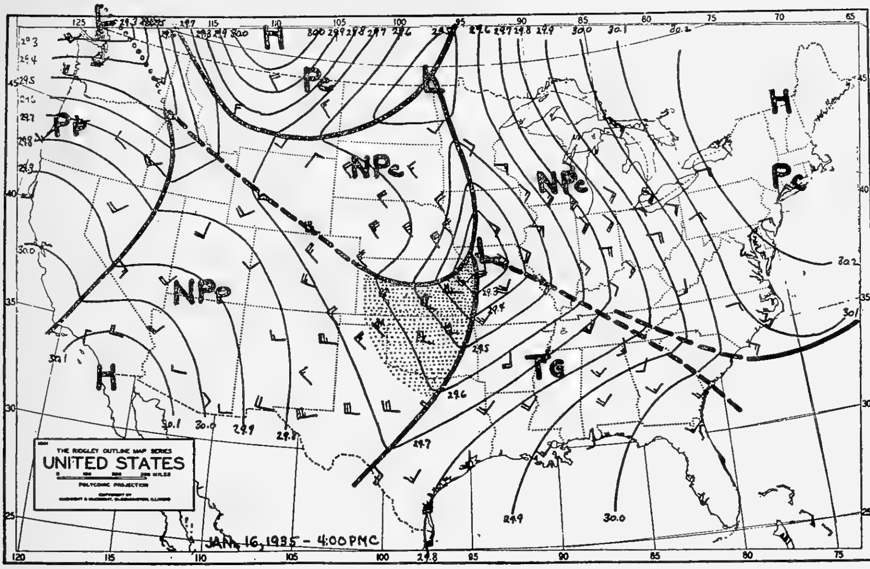


FIG. 3. WEATHER MAP, 4 P. M., JANUARY 16, 1935.

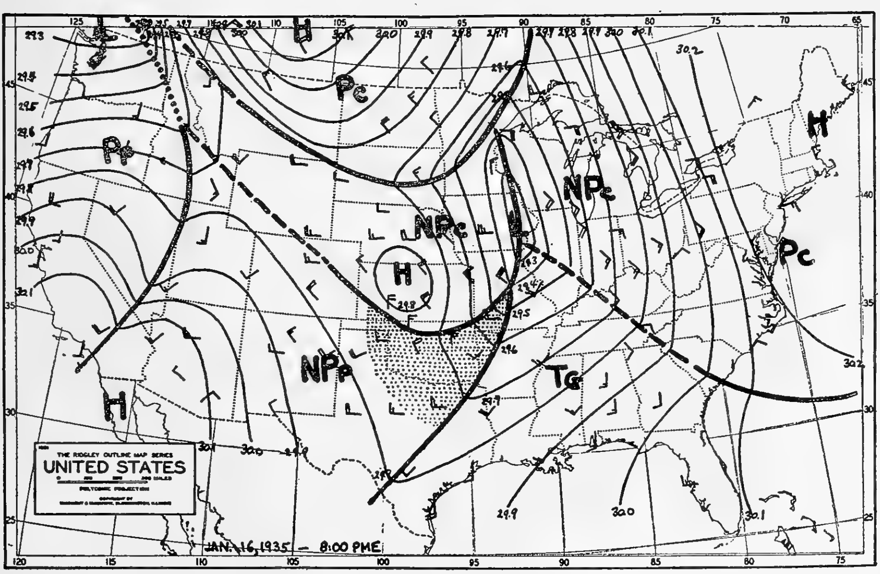
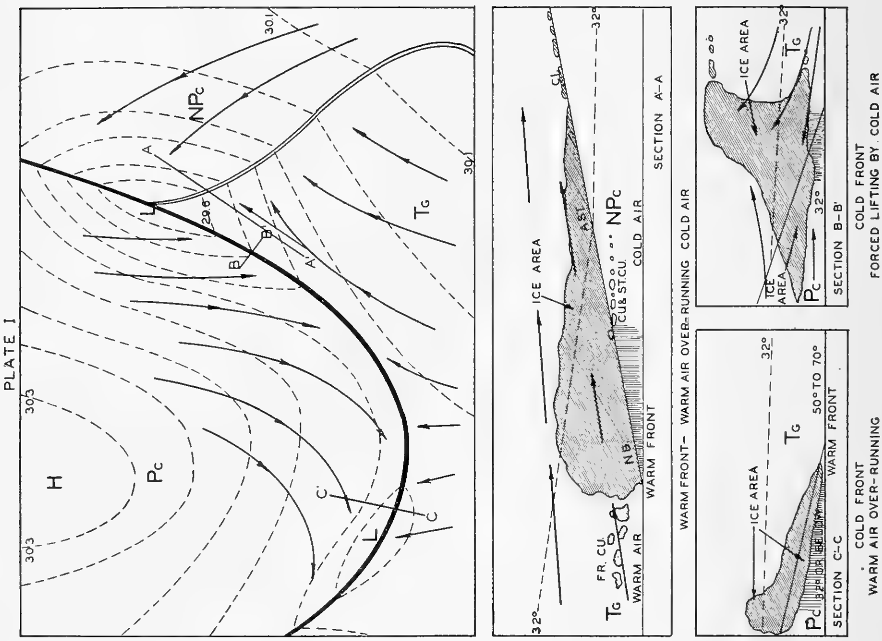
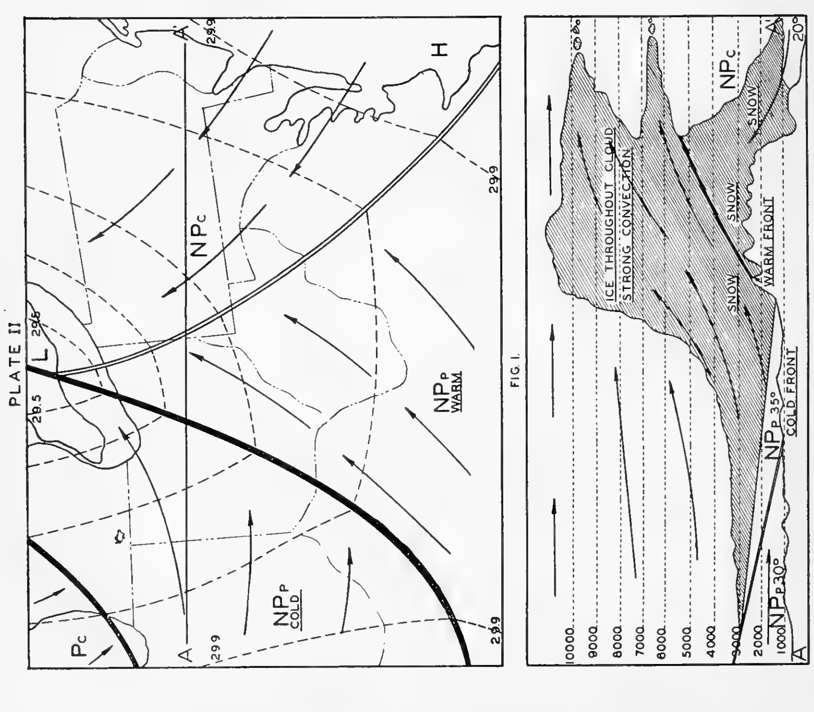


FIG. 4. WEATHER MAP, 8 P. M., JANUARY 16, 1935.

Maps and Cross Sections of Fronts: Aircraft Icing
(From E. J. Minser, BULLETIN, May 1935, p. 132)





WARM FRONT – WARM AIR OVER-RUNNING COLD AIR – MOUNTAIN BARRIER ACCELERATING CONVECTION. ICE FORMING IN ALL PORTIONS OF CLOUD, MOST INTENSE IMMEDIATELY OVER WESTERN RIDGES WHERE CONVECTION IS STRONGEST.

A Succession of Polar Air Masses: Weather Maps and Cross-Sections for Nov. 30-Dec. 2, 1938

The charts below are from an article by George and Elliot in the *Bull. Amer. Met. Soc.* for March 1939. The period was featured by an outbreak of PC air over the Great Lakes region, and a series of rapidly moving invasions of NPP air across the western states. NPC is stagnating in the Southeast, and a tongue of TA begins to move over Texas from the Gulf. The cross-sections from Oakland to Washington show the considerable depth of the PP compared to the PC air and the tendency for warm-front-type occlusions and upper cold-fronts to form from PP invasions. (Cold fronts are drawn solid black, warm fronts double lined, occluded fronts dashed, upper fronts dotted; temperature in °F and winds are entered for a few stations).—*R. G. S.*

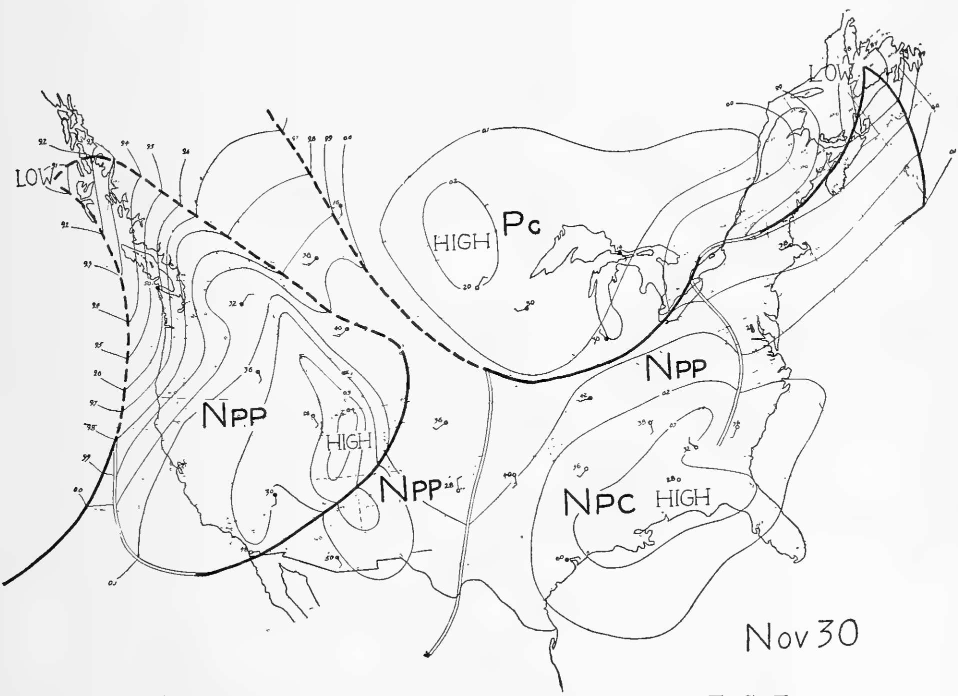


FIG. 1. Synoptic Chart, Nov. 30, 1938, 7:10 a.m., E. S. T.

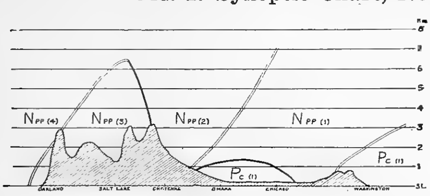


FIG. 2. Cross-section, Oakland to Washington, 4 a.m., E. S. T., Nov. 30, 1938.

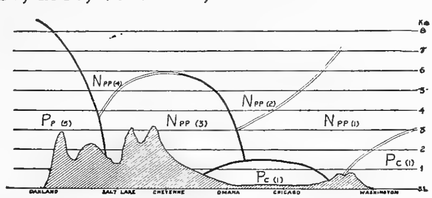


FIG. 4. Cross-section, 4:00 a.m., Dec. 1, 1938.

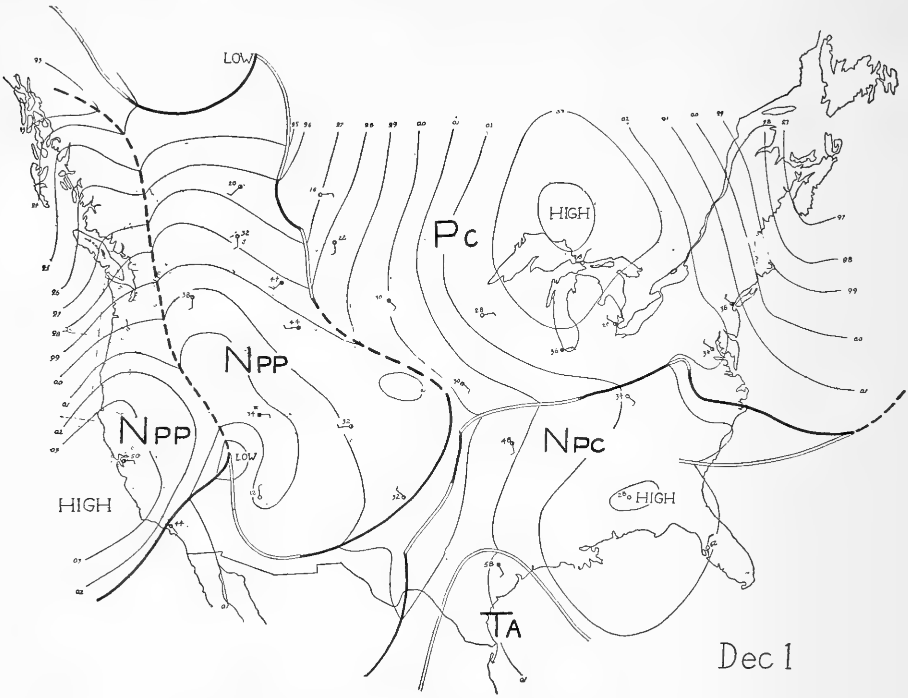


FIG. 3. Synoptic Chart for Dec. 1, 1938, 7:10 a.m., E. S. T.

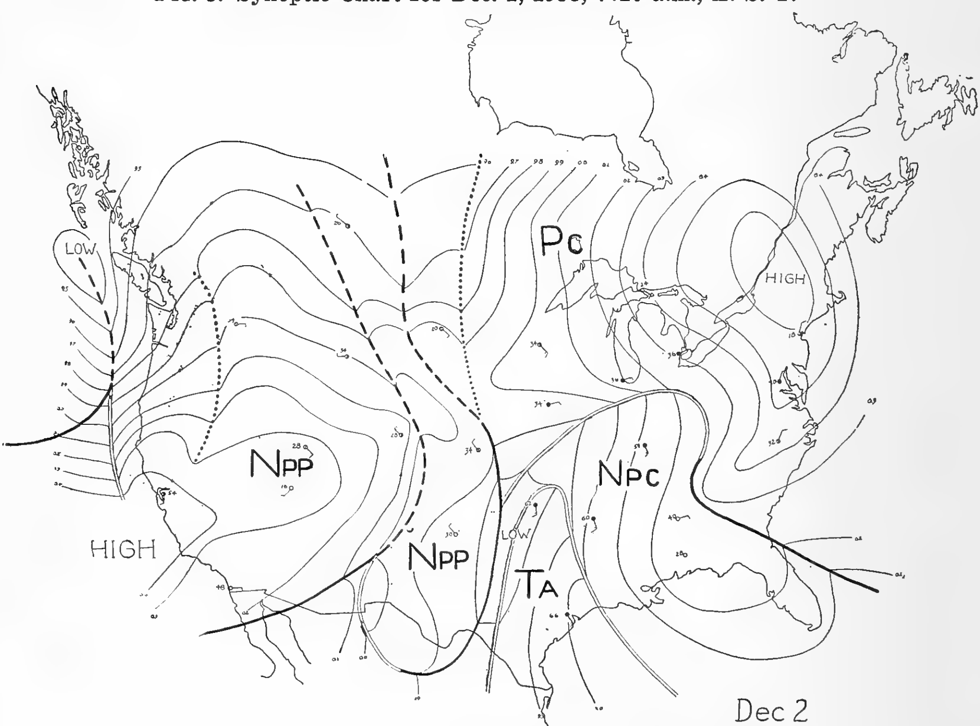


FIG. 5. Synoptic Chart for 7:10 a.m., E. S. T., Dec. 2, 1938.

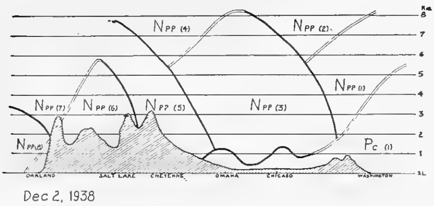
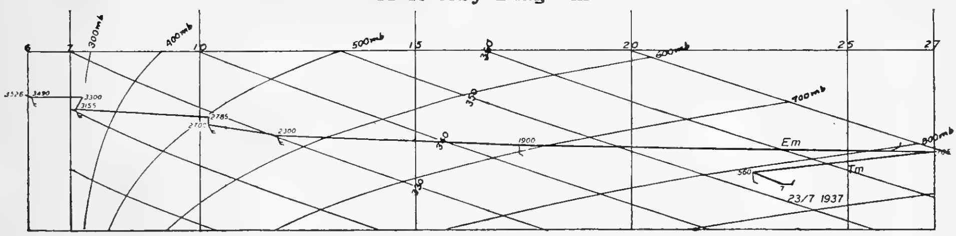


FIG. 6. Cross-section for 4:00 a.m., E. S. T., Dec. 2, 1938.

A Rossby Diagram



A ROSSBY DIAGRAM OF AN AIRPLANE SOUNDING AT NANKING, CHINA, AUG. 23, 1937. There is a shallow layer of TM air at the surface with a sharp front at 706 meters, above which Equatorial marine air exists to the top of the sounding. This sounding was made just after a typhoon passed inland over Nanking, the warmer EM being imported *aloft* by the circulation of the storm. (From a paper by Tu, *Bull. Amer. Met. Soc.*, March, 1939.)

Synoptic Charts Showing Winter Cyclones

(Figs. 1, 2, 6-13 from: Dorsey, *Bull. Amer. Met. Soc.*, Oct. 1938.)

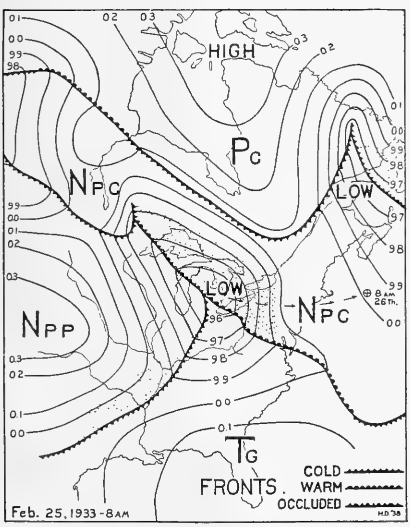


FIG. 1. Deep cold air over New England was augmented by fresh PC the next day. The warm and occluded fronts in the low over Michigan are producing snow (stippled area) over a wide region and caused heavy snow in New England later the same day as it moved eastward.

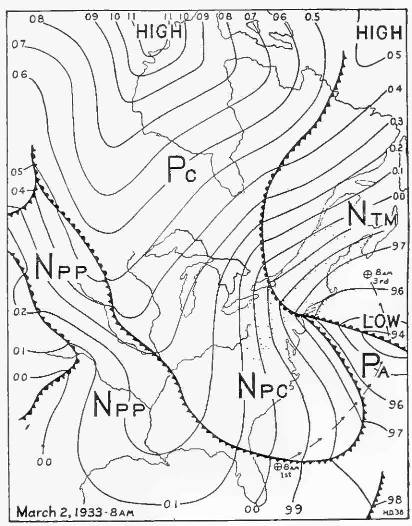


FIG. 2. The low off the east coast was formed by amalgamation of two weaker lows the day before. This storm caused N W gales and heavy snow on the seaboard. In this and Figs 2-4 the effect of fresh PC outbreaks in sweeping up old occlusions ahead of them is evident.

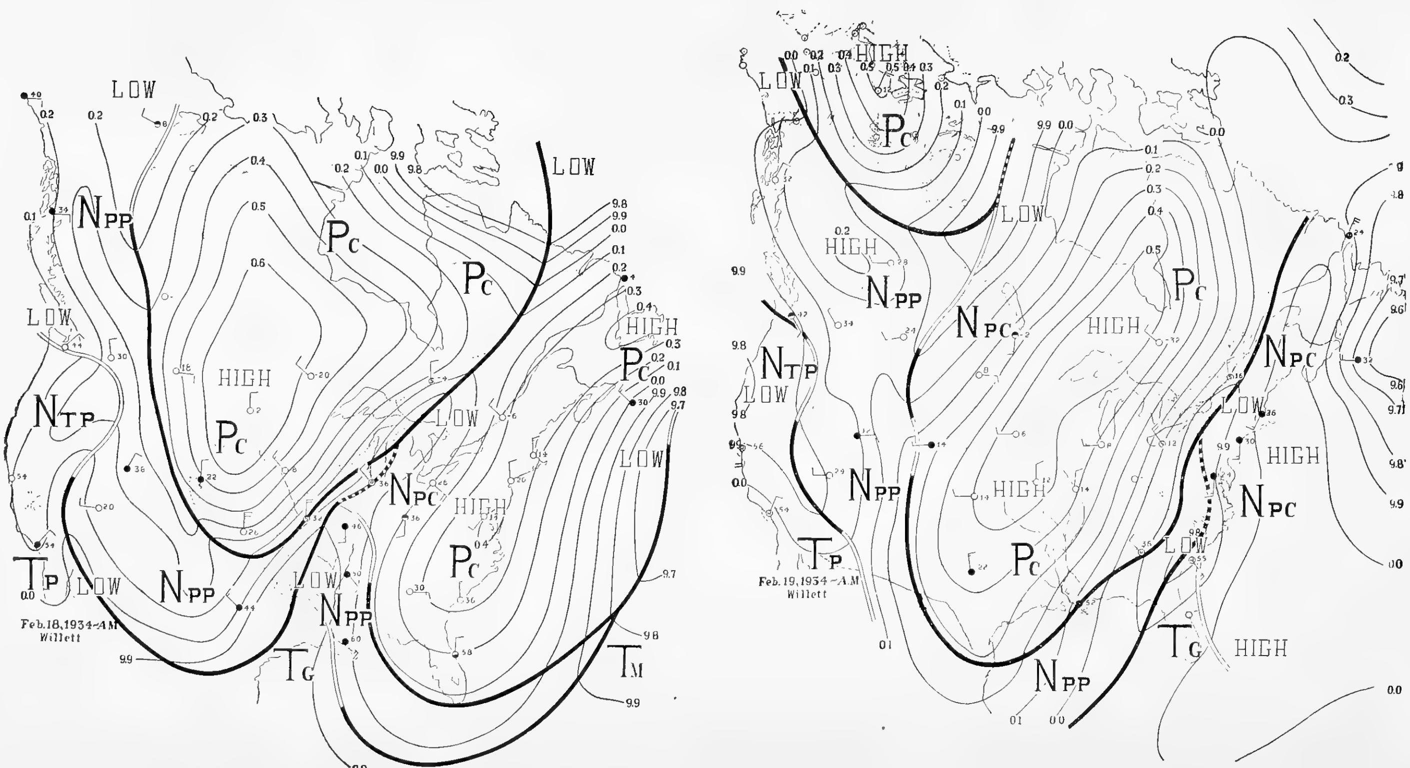


FIG. 3. A series of maps copied from the M. I. T. file analyzed by Prof. Willett; these were discussed briefly by C. H. Pierce in the *Bull. A. M. S.*, 1934, pp. 61-78. The data for most of the stations has been omitted in order to avoid confusion.

(Temperatures are shown by figures east of station. Wind velocities shown by the barbs at end of direction arrows,—one short barb for each Beaufort number. Double parallel lines represent the warm fronts, heavy solid lines represent cold fronts, and broken lines occluded fronts. Stippled area shows the precipitation area at time of observation. The following Fig. 4 is the same.)

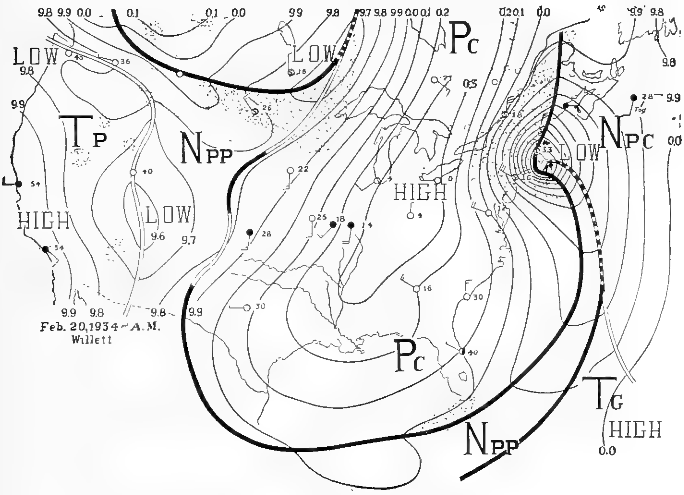


FIG. 4 shows the sudden intensification of the coastal storm and the subsequent movement of the Pc air mass shown in Fig. 3.

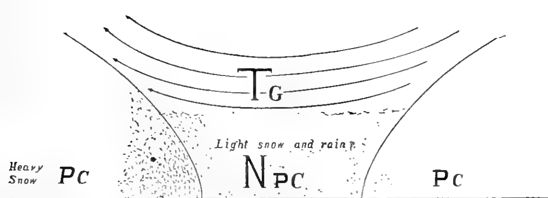


FIG. 5 is a W-E cross-section of the air masses over southeastern New England and just north of the cyclone center at 8.00 a.m. the 20th, showing the occlusion of TG air causing heavy snowfall around New Haven due to lifting of the TG, but only light snow at Fall River where the TG is not being lifted much (see Fig. 4).

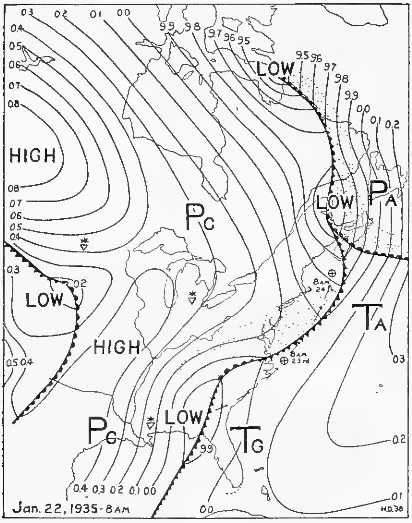


FIG. 6. The N-S alignment of the Pc cold front results when cold air passes rapidly south to the Gulf while the tropical air is not displaced eastward very far. A wave-disturbance has started in the southeast and an older one is occluding over New Brunswick. The western low joined with the southeastern one the next day and then moved NE-ward with the Tm air spreading westward aloft over the Pc causing very heavy snow and N-ly gales in New England. (Legend for this and Figs 7-13 same as in Fig. 1.)

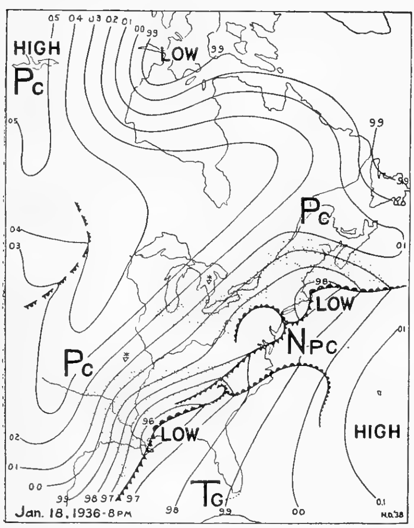


FIG. 7.

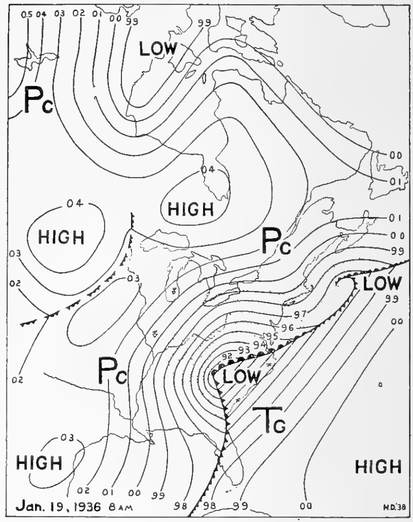


FIG. 8.

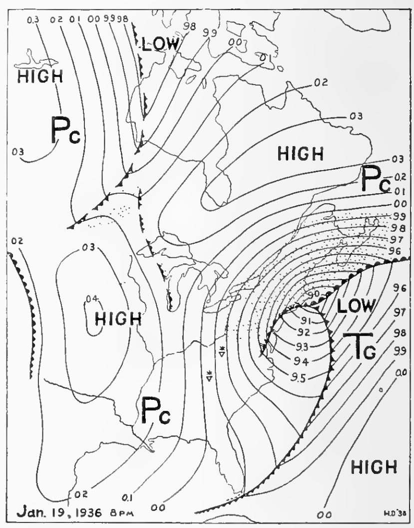


FIG. 9.

FIGS. 7, 8. and 9. A succession of weak occlusions and young wave disturbances along a quasistationary cold front, the last wave deepening and occluding into an intense cyclone with heavy snow and gales over the northeast.

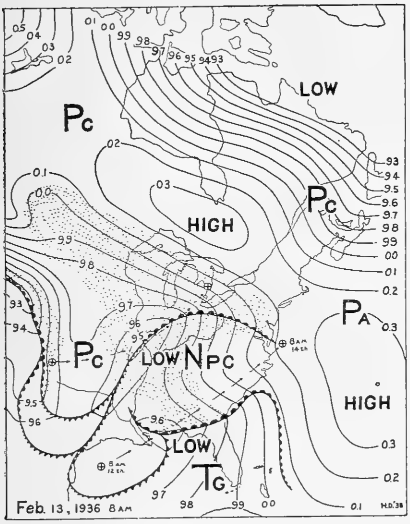


FIG. 10. A somewhat complicated situation accompanied by widespread precipitation. The three occluded lows resulted from a series of minor wave disturbances that came in from the Pacific. The southern low deepened further and occluded off Hatteras, causing heavy snow in Boston on the 14th.

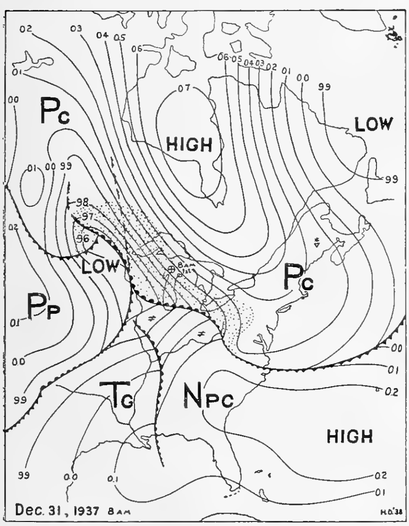


FIG. 11. A large occluding Pacific low moving eastward along the western Pc front eventually drew TG into its circulation causing intensification and a new wave which was occluding the next day over New England. Note the upper cold front (warm-front occlusion) in the west where the PP-Pc occluded front (NPc occluded) is moving aloft over the Pc. The TG will soon enter into this upper occlusion by rising over the NPc and displacing it.

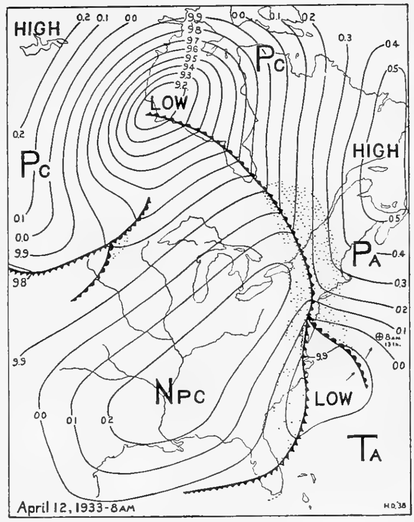


FIG. 12. The deep occlusion in Canada had been formed by TG intensifying an older occlusion that had advanced from the Pacific along a Pc front to near Omaha. The Pc spread south behind the low to California and by the 12th was as shown here modified to NPc with a fresh Pc outbreak in the NW. A secondary low, in the southeast, formed at the base of the occlusion. It moved NE between rising pressure at Bermuda and the PA current flowing down from the N, occluding off the New England coast where it caused Nly gales and snow. The northern low is one of the spring type that sometimes develop into deep cold vortices without fronts—"dynamic lows".

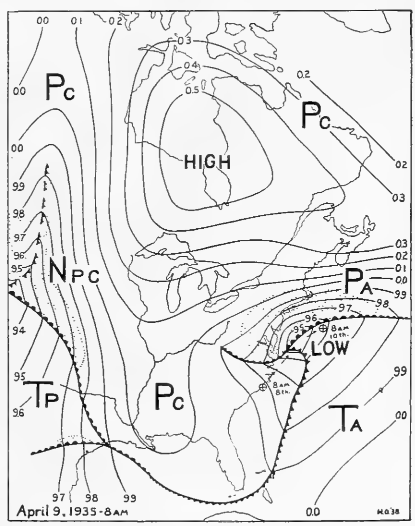


FIG. 13. This shows a typical situation for PA invasion of the northeastern seaboard, induced by a slow moving high to the north and an occluding wave disturbance from the southeast. As the low moves NE, the PA may continue for days on the seaboard.

X. ISENTROPIC ANALYSIS *

JEROME NAMIAS

INTRODUCTION

IN AIR MASS and frontal analysis use is made of "indirect aerology" —a technique of deducing chiefly from observations of clouds and hydrometeors the nature and structure of the atmosphere above the ground. This method supplies a good deal of information essential to weather analysis and forecasting, and is especially valuable where direct observations of the upper air are lacking. Where a fairly dense network of upper-air soundings is available, however, indirect aerology naturally must give way to the consideration of *observed* conditions. In the United States we are fortunate in being able to make use each day of some thirty radiosonde observations, and it appears that this number will increase from time to time. Indirect aerology is thus being forced more and more into the background; but there arises the problem of how to use effectively these upper-air data in the daily routine of analysis and forecasting. The current practice is all too often to carry out an analysis of surface weather maps, afterwards using the upper-air data merely as a check. This form of treatment, it hardly need be stated, rarely leads to the development of new and directly usable ideas for interpreting the soundings.

Appreciable progress in the use of upper-air data has recently been made by Rossby and his colleagues [3][1] at the Massachusetts Institute of Technology; their technique, called *isentropic analysis,* goes further than merely supplementing the surface analysis of air masses and fronts; it brings to light much entirely new knowledge of the physical processes at work in the atmosphere. Isentropic analysis has already become an integral part of the modern forecaster's technique, and is now widely used in the United States.

§ 1. BASIS FOR THE ANALYSIS

In Articles II to IV we discussed the need in synoptic meteorology for conservative elements by means of which parcels of air may be identified from day to day. It was pointed out that two of the most conservative of these elements are potential temperature and mixing ratio or specific humidity, both of which do not change during adiabatic processes as long as the air remains unsaturated, and as long as turbulent redistribution of heat and moisture may be neglected. These two quantities are used as coördinates of the Rossby-diagram [1][1], and if the points of an aerological sounding are plotted on such a diagram we obtain the "characteristic" curve. This curve is unaltered during any adiabatic process involving the same particles, for in this case the individual points of the curve stay fixed. Thus the characteristic curve may be used to identify vertical air columns as they move over the surface of the earth. Under the assumption made above, each potential temperature in a series of characteristic curves obtained at different stages in the history of a moving air column will then be characterized by a practically unchang-

[1]See references at end of this chapter.

*Reproduced, with changes, from the MS of the chapter under the same title which appears in the book "Weather Analysis and Forecasting," by kind permission of Prof. Sverre Peterssen and the McGraw-Hill Book Company.

ing mixing ratio. We may therefore choose some particular surface of potential temperature and use the mixing ratio on this surface as an identifying element by means of which parcels of adiabatically moving air having the chosen potential temperature may be traced from day to day. The equation (see Article VIII)

$$S = C_p \log \theta + \text{constant}$$

expresses the fact that a surface of constant potential temperature is also a surface of constant entropy, and hereafter we shall speak of it as an *isentropic surface.* Since the atmosphere is normally stable, the potential temperature increases steadily with elevation. We may then consider the atmosphere as consisting of an infinite number of thin isentropic sheets limited by the surfaces θ = constant, $\theta + d\theta$ = constant, etc., etc.

The above considerations, in part, led Rossby [2, 3] to suggest that upper-air charts be drawn along isentropic surfaces, rather than along constant levels where the comparison of elements from point to point is at times misleading because of ascending and descending motion of the air through these level surfaces. Sir Napier Shaw [4] had suggested some years ago that weather maps be drawn along isentropic surfaces, and had actually constructed a few such charts containing isotherms. He pointed out that isotherms on an isentropic surface were also lines of constant density and constant pressure. But Shaw did not make use of the mixing ratio (w) as a second identifying element to be used on isentropic charts. For numerous reasons, probably the main one being the lack of aerological data at the time, Shaw's suggestion was not put into practice, and the method of analyzing conditions along isentropic surfaces lay dormant until revived by Rossby several years later. But other considerations, the fruit of later studies, helped to indicate that isentropic surfaces should be used.

In a study of the Gulf Stream, Rossby [5] found much evidence of large-scale cross-current mixing between the Gulf Stream water and its environment, and that this mixing takes place along surfaces of constant density[2]. In the atmosphere, where compressibility must be taken into account, it is readily seen that this type of mixing must operate chiefly along isentropic surfaces. In figure 1, for example, we have a normal atmospheric stratification in which the po-

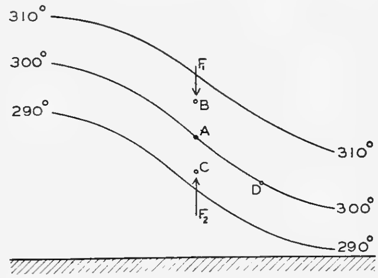


FIG. 1. FORCES RESISTING DISPLACEMENTS FROM ISENTROPIC SHEETS.

tential temperature increases with elevation. A parcel of air originally resting at A, if displaced to the point B, will be subjected to a downward force F_1, since it finds itself colder than its environment (see Art. II). Similarly, if the parcel is displaced to C a force F_2 resists the displacement. These restoring forces are proportional to the vertical gradient of potential temperature. But if A is displaced to D, that is, along an isentropic surface, there is no resisting force, since at each point the particle has the same temperature and density as its environment. Thus lateral mixing in the atmosphere must take place chiefly along the surfaces of constant potential temperature (isentropes). In saturated air the mixing operates

[2]Strictly speaking, surfaces of constant potential density.

chiefly along surfaces of constant equivalent-potential temperature. There is also some evidence in support of the view that the intensity of vertical mixing decreases and the intensity of lateral mixing increases as the vertical stability increases. This latter conclusion is known as Parr's principle[6].

The importance of isentropic mixing in the atmosphere lies in the fact that if it is of appreciable magnitude, this mixing must lead to sizable shearing stresses operating across planes normal to the isentropic surfaces, whenever there are variations in wind velocity in a broad current flowing along the isentropic surface. If a westerly current flowing along an isentropic surface is of such character that the velocities are larger to the north and diminish to the south, shearing stresses will tend to speed up the westerlies to the south and retard the eastward flow along the northern edge of the current. In other words, the shearing stresses tend to distribute momentum uniformly over all filaments of the current. As a result of these stresses, frictional volume forces are set up which act in the direction of the axis of the current, retarding in the regions of velocity maxima, accelerating in the regions of velocity minima. Under steady state conditions these axial forces must be balanced by Coriolis forces associated with slight motions normal to the current axes. Thus, as pointed out by Ekman [7], in the Northern Hemisphere an accelerating force F, per unit mass, acting eastward, produces a southward motion whose velocity is

$$v = \frac{F}{2\,\omega \sin \phi} .$$

Fig. 2 shows diagrammatically the accelerating and retarding forces F_a and F_r, respectively, which operate on

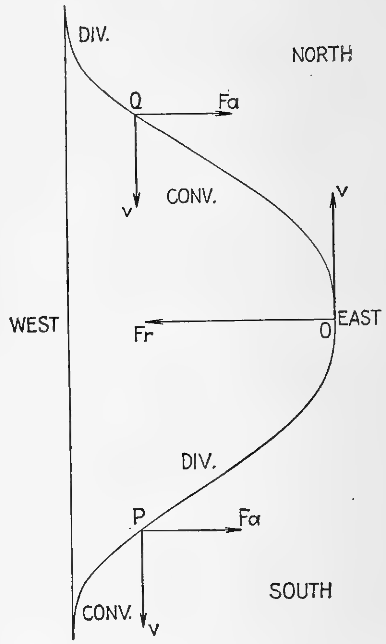


Fig. 2.—Accelerating and retarding forces operating on an isentropic current profile in which velocity varies in neighboring filaments.

account of lateral mixing within a broad westerly current whose velocity profile is indicated by the curve. The motions which result from the Coriolis force are given by the arrows marked v.

It will be seen from Fig. 2 that air motions created by shearing stresses may result in regions of convergence and divergence. For example, at P air is being flung to the south, while at O it is being flung to the North, so that between these two filaments divergence must set in. Similarly, convergence must set in in the region between O and Q. At the ground in regions above which convergence of this nature is taking place at all levels the pressure rises, while below regions of divergence the surface pressure falls. Far to the right of the stream the lateral shearing stresses will also produce convergence, for to the extreme south the velocities are hardly accelerated at all. Therefore, it is highly probable that *lateral shearing stresses may be important in producing cross-isobar wind components,* and thus also pressure variations which are of a purely dynamic rather than thermal-advective nature. There has been accumulated an increasing mass of evidence, both on theoretical and observational grounds, that the magnitude of these lateral shearing stresses is sufficient to account for such pressure variations. Thus by equating isentropic shearing stress within sloping parallel isentropic surfaces which intersect the ground at a normal angle and stress due to ground friction (and assigning reasonable values of lapse rate and wind velocity in temperate latitudes) Rossby [8] obtained a lateral stress of 167 dynes/cm^2 corresponding to an isentropic shear of 2.5×10^{-5} sec^{-1} and an isentropic eddy viscosity of 6.7×10^6 grams/cm/sec. These values agree fairly well in order of magnitude with those obtained from a study of the diffusion of water vapor actually observed in isentropic charts (Grimminger [9]); moreover, they lie between those found by Richardson and Proctor for diffusion over small distances and those obtained by Defant by considering cyclones and anticyclones as large-scale turbulent elements in the general circulation. The analysis of isentropic charts has shown that eddies of appreciably smaller size than extra-tropical cyclones and anticyclones occur almost daily in the isentropic flow patterns.

There is being gathered an increasing mass of observational evidence to substantiate the theory summarized above[3]. In the following pages we shall briefly discuss the practical results of studies of some of this material.

Before proceeding with this discussion it is well to repeat that the isentropic analysis suggests itself as a practical tool in synoptic meteorology for the following two reasons:

1. It provides a method of identifying and following large-scale moist and dry currents and of anticipating their subsequent thermodynamic modifications.

2. It provides a method for taking into consideration lateral shearing stresses and their hydrodynamical effects on the prevailing flow pattern.

§ 2. Plotting Routine

The mechanical operation of preparing upper-air data in a form suitable for isentropic analysis is quite simple. We require for the analysis two sets of upper-air charts: isentropic charts, for which an ordinary geographical base map suffices, and atmospheric cross sections which represent vertical planes cutting through the atmosphere along desired lines. The choice of some convenient isentropic surface or surfaces for analysis must then be made. The deciding factor in this regard is the general elevation of the surface, which again depends upon the general temperature distribution. Thus the isentropic surface $\theta = 290°$ which has been found useful in the United States during the winter months is far too low during the summer months. In general, the isentropic sheet finally chosen should be high enough to be above the layer influenced by surface friction, and at the same time, low enough to show a large range of specific-humidity values. It is obvious that over a large area of diverse topographical and climatic features these two criteria

[3]A complete quantitative treatment of the underlying theory is at present in preparation by C.-G. Rossby and his colleagues.

cannot be completely satisfied. However, it is in general possible to find an isentropic surface that satisfies these conditions well enough for a reasonably sound analysis.

In North America suitable values of potential temperature for the individual seasons are:

Season	*Potential Temperature*
Winter	290°-295°A
Spring	295°-300°A
Summer	310°-315°A
Fall	300°-305°A

The abrupt increase from spring to summer values is due to the normally rapid increase of free air temperatures as summer convection sets in.

During periods of unusual weather it is sometimes necessary to change to a different surface for a few days, and then to return to the normal for the season. It should be pointed out, however, that such changes carry with them a certain loss of continuity—and it cannot be emphasized too strongly that continuity is the primary requirement of isentropic analysis. For this reason it is most advantageous to follow from day to day the flow patterns in a given isentropic surface, and when an abnormal period suggests change of surface, to construct an additional set of charts for a more representative surface for this particular period.

The elements plotted along the isentropic surface are the mixing ratio, saturation mixing ratio, and atmospheric pressure[4]. The appropriate values for plotting on the isentropic chart are readily extracted from aerological soundings with the help of some convenient thermodynamic diagram (see Article III). The pilot-balloon wind observations are then entered along the isentropic surface for those stations where the height (or pressure) of the isentropic surface is already evaluated. In addition to the above data, it is helpful to indicate the form and motion of clouds as well as hydrometeors observed at the aerological sounding stations together with the levels in which these phenomena are reported.

The atmospheric cross-section diagrams now in use in United States have pressure (on a logarithmic scale, but $p^{0.288}$ scale is being introduced) as ordinate and horizontal distance as abscissa. The values of mixing ratio and potential temperature are plotted at the significant levels for each station. If time permits it is convenient to plot also the relative humidity and the temperature. By constructing lines of constant moisture and constant potential temperature in the cross sections we obtain a picture of the moisture and temperature distribution along the vertical plane represented by the cross section, as well as a view of the moisture pattern in the isentropic surfaces. Pilot-balloon wind observations, clouds and hydrometeors are also plotted in the cross sections.

§ 3. Technique of Analysis

In order to enter all available pilot-balloon winds on the isentropic surface it is necessary to sketch a set of lines of constant pressure. We shall refer to these as contour lines. Usually it suffices to draw such lines for each 50th millibar. The number of aerological soundings is in general insufficient for drawing the contour lines in a purely mechanical fashion. Contour lines may be drawn more accurately by taking into consideration the guiding factors mentioned below. It is, indeed, necessary to make use of these aids when one wishes to extend the analysis into regions in which the data are sparse.

[4] *Byers* [10] has developed the use of condensation pressure instead of mixing ratio, and that method has been adopted in the U. S. Weather Bureau.

The particular isentropic surface or surfaces to be analyzed may be sketched immediately in the cross sections, and the positions of troughs or ridges may thus be determined and indicated on the isentropic map.* In sketching the isentropes on the cross sections, one should bear in mind that in the case of adiabatic lapse rates, (i.e., $d\,\theta/dz = 0$) the isentropes run vertically. In general, one can determine the lapse rate directly from the isentropes by making use of the following simple derivation. As an approximation we may write:

$$\theta = T + z \qquad ,$$

where T is the temperature and z the height above sea level expressed in hundreds of meters. Differentiating with respect to elevation we obtain:

$$\frac{\partial\,\theta}{\partial\,z} = \frac{\partial T}{\partial\,z} + 1, \text{ or } \frac{\partial\,T}{\partial\,z} = \frac{\partial\,\theta}{\partial\,z} - 1\,.$$

If one chooses for dz a unit distance, say 1000 m, then $\frac{\partial\,T}{\partial\,z}$, is the lapse rate, determined by simply noting the difference of potential temperature in the vertical range; or the distance between two successive isentropes may be estimated, and $\frac{\partial\,T}{\partial\,z}$ quickly determined. This method is especially helpful when lapse-rate diagrams are not at hand.

The domes and ridges in the isentropic surfaces are found in regions occupied by cold air masses, and the troughs in the isentropic surfaces are found in the warm air masses. To the extent that the surface pressure changes are due to advection of cold and warm masses, the slope of the isentropic surfaces will be greatest above the isallobaric maxima.

Fronts on the surface weather map also offer a great deal of information about the structure and pattern of the contour lines. Regions in the vicinity of well marked fronts are characterized by a crowding of the contour lines. Since cold fronts are generally much sharper and steeper than warm fronts, this steep gradient of contour lines on the isentropic chart is usually at its maximum just behind cold fronts. Moreover, as is to be expected, the contour lines usually run parallel to the surface fronts. Experience shows that most well-defined fronts are characterized by constant, or almost constant, potential temperature, so that the frontal surfaces have a marked tendency to coincide with the isentropic surfaces. Moreover, since a frontal surface is characterized by stable stratification, $\frac{\partial\,\theta}{\partial\,z}$ would have a maximum within the transition zone. Since lateral mixing occurs mainly along isentropic surfaces, it follows that frontal surfaces which are parallel to isentropic surfaces will not be destroyed by mixing. On the other hand, frontal surfaces which intersect the isentropic surfaces will dissolve on account of lateral mixing, unless the front is situated in a field of motion which is pronouncedly frontogenetical. When the front is associated with large areas of precipitation, the process is no longer isentropic, and lateral mixing does not take place along isentropes, but along surfaces of constant equivalent-potential temperature or surfaces of constant saturated potential wet bulb temperature. In such cases the isentropic surfaces will cross the frontal surfaces. At cold fronts, however, the area of precipitation is normally relatively narrow so that, in most cases, cold fronts lie along isentropic surfaces above the frictional layer. The warmest air is normally found just ahead of the sur-

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*Prof. Spilhaus has suggested a more refined technique for drawing the isentropes (*Bulletin Amer. Met. Soc.*, June, 1940).

face cold front. Thus the trough in the contour lines is usually found slightly in advance of the cold front, with a steep upward slope of the isentropic surface in the rear of the cold front.

The parallelism of contour lines with well-marked cold fronts many times enables one to construct height lines in regions where upper-air observations are sparse. The application of this principle to frontal-wave disturbances is apparent, for here there must be a wavelike pattern in the contour lines roughly parallel to the frontal waves in the synoptic surface chart.

Sharp warm fronts show up in the contour-line pattern in much the same manner as do cold fronts, the gradient increasing abruptly at the front while the lines remain fairly parallel to the front. There are, however, many warm fronts on the surface weather maps which are not associated with this simple contour-line pattern above. Moreover, the surface maps frequently indicate an homogeneous air mass in the warm sector, while the isentropic charts show conclusively that the warm air is far from homogeneous aloft, but rather characterized by troughs in the contour lines suggestive of fronts. Some time *after* their appearance in the contour-line pattern, these troughs may appear on surface charts as regions of frontogenesis, suggesting that such newly formed fronts are a *result* rather than a *cause* of the processes in the upper air. It thus appears that many fronts on the surface weather map are induced by action taking place first in the upper air and later showing up at the ground. This, obviously, means that a frontogenetical wind-field develops aloft and gradually extends downwards.

The above ideas are illustrated in Fig. 3:—The light broken lines labelled z, $z + 1$, etc., are contour lines of an isentropic surface. H indicates regions where the isentropic surface is high, and L regions where it is low. Fig. 3a represents the normal topography, while Fig. 3b shows a type of topography which is associated with frontogenesis within the warm air mass subsequent to the appearance of the trough in the contour lines. This latter case is presumably associated with frictionally driven anticyclonic eddies which we shall treat in more detail later on.

Finally, we may use the upper-air wind observations as entered directly for the observed heights of the isentropic surface over aerological sounding stations as a guide to the configuration of contour lines. In cases where the pressure distribution aloft is largely determined by the distribution of temperature, the winds blow nearly parallel to the contour lines so that domes or ridges are generally to the left of the current flow.

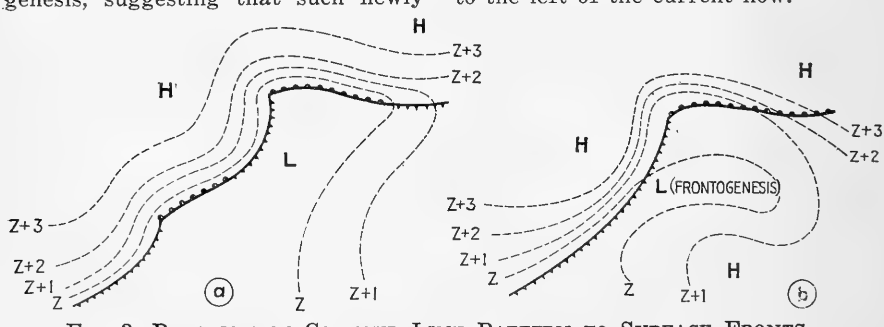

Fig. 3. Relation of Contour Line Pattern to Surface Fronts.

§ 4. Isentropic Flow Patterns

An inspection of the distribution of moisture along an isentropic surface covering a sufficiently large area immediately brings to light the fact that there are regions of high and of low moisture concentration. If the network of aerological stations were sufficiently dense, it would be possible to draw mechanically a series of lines, each denoting a given mixing ratio. In this manner one could locate sources of moisture, or regions of injection of moist tongues, and likewise the regions from which dry air is supplied. Moreover, these currents could be followed in a continuous fashion from day to day as they travel along the isentropic surfaces. Unfortunately the distribution of aerological stations in any part of the world is much too sparse to permit such a mechanical delineation of moist and dry currents, and for this reason it becomes necessary to develop models and indirect clues by means of which characteristic "flow patterns" may be drawn which approach the real solution.

The topic of the source of our moist and dry currents is reserved for a later section of this chapter. For the present we shall treat the fundamental flow patterns which thus far have established themselves in the daily isentropic analysis. Once these models are recognized on the daily isentropic chart, the analysis of the moisture lines becomes appreciably simplified.

The patterns of the large-scale motions in the atmosphere appear to be controlled in large part by the rotation of the earth. If we consider a fluid chain of particles located in and moving with an isentropic surface, it follows from Bjerknes' circulation theorem that the total *absolute* circulation of this chain remains constant as it moves from latitude to latitude. The absolute circulation (C_a) is equal to the sum of the circulation of the fluid chain relative to the earth (C_r) and the absolute circulation obtained by the chain if it momentarily were fixed to the earth (C_ω), or:

$$C_a = C_r + C_\omega \quad .$$

Positive values of C_a and C_r indicate circulation with cyclonic sense, negative values anticyclonic sense. It can be shown that C_ω is given by

$$C_\omega = 2\,\omega\,\Sigma \quad ,$$

where Σ is the area enclosed by the projection of the fluid chain on the equatorial plane. Thus, if an originally stationary isentropic fluid chain moves northwards without change of the horizontal area it encloses, its equatorial projection increases, and since $(C_r + 2\,\omega\,\Sigma)$ shall remain constant, C_r must decrease; hence, the chain will gain an increasing amount of anticyclonic circulation which adds to the circulation originally possessed by the chain. Similarly, a southward moving system tends to develop a cyclonic circulation. Cyclonic circulation normally expresses itself as a cyclonic curvature of flow, anticyclonic circulation as an anticyclonic curvature of flow. Thus the polar currents of cold dry air coming out of the north are generally cyclonically curved, while the warm, moist flows from a southerly direction have anticyclonic curvatures. These large-scale flows are the fast-moving streams and generally dominate the flow patterns observed on the daily isentropic charts. The axes of such streams may be delineated on the isentropic charts as curved lines along which the wind velocities are a maximum. Once these current axes are determined, they serve as a framework for isentropic flow patterns.

Once an air mass is set in motion, certain adjustments of the pressure field within and surrounding the current must take place more or less as they do in the case of the wake stream, investigated by Tollmien [11]. Thus if momentum is injected into a region where no horizontal pressure gradients previously exist, Rossby [12] has shown that there will be a banking of the current so that pressure rises along the right edge of the stream and falls along the left edge. This effect is due to the action of an initially unbalanced Coriolis force which attempts to create a new pressure distribution in order to balance the motion. There is thus a tendency for a mutual adjustment of pressure and velocity distributions. These adjustments are affected whenever atmospheric flow becomes out of balance with its pressure gradient.

The results of Rossby's theory which are of immediate practical value in isentropic analysis are:

1. Fast-moving streams tend to suffer a "banking" process so that a ridge of pressure tends to build up to the right (in the Northern Hemisphere) and a trough to the left of the stream.

2. The action of lateral shearing stresses operating on a current in which the velocity varies (as in Fig. 2) produces super-gradient winds along the boundaries, and thereby causes air to be flung across the isobars from lower to higher pressure, and this tends to build up higher pressure along the right edge of the stream.

3. The shear zones on either side of fast-moving currents are dynamically unstable (as shown by Pekeris [13]) and tend to break the current into eddies having the vorticity of the original current profile. Thus, anticyclonic eddies are formed to the right of fast-moving streams, cyclonic eddies to the left.

The development of an eddy affects the moisture distribution and creates a distinct pattern of moisture lines along an isentropic surface. In fig. 4a, for example, we have a zonal distribution of moisture in an isentropic surface in middle latitudes. The isen-

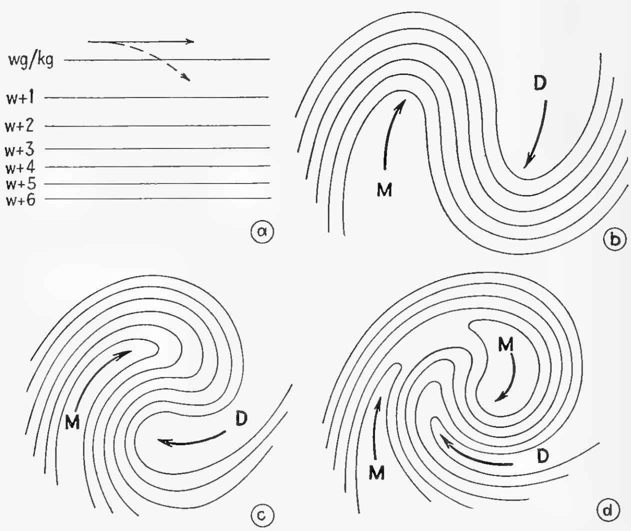

FIG. 4.—SUCCESSIVE STAGES, (a), (b), (c), (d), IN THE DEVELOPMENT OF AN ANTICYCLONIC EDDY AS REVEALED RY THE MOISTURE LINES.

tropic surface itself is higher to the north and tilts southward, so that it may be 5000 m above sea level in the north and only 2000 m high in the south. Let us now superimpose upon this zonal state a velocity field as indicated by the lower half of fig. 2. Then air to the south is accelerated through lateral shear, air is piled up to the right of the accelerated stream, and the motion takes on a circulatory pattern indicated by the broken arrow. Since the moisture is carried along by the air and the motion is adiabatic, successive patterns of moisture lines indicated in figs. 4b to d are soon developed.

Once the moist tongue of an eddy is sketched in on an isentropic chart it is possible, with the help of the cross sections, to apply tests for the existence of the tongue in the specific area chosen. The basis for this test lies in the empirical fact that significant moist tongues spread out aloft and if a moist tongue is present on an isentropic surface *between* aerological stations which are not more than 400 to 600 kilometers apart, it generally shows up in the cross section as an inversion of mixing ratio (or at least a minimum in the vertical gradient of moisture) at one or both of the stations. If there are no inversions of moisture above the chosen isentropic surface at either of the stations in the cross section, and if a moist tongue is assumed to lie between stations, the tongue must be drawn as a narrow vertical filament of moist air—a highly improbable condition, and certainly one that is rarely observed. This test is perhaps more clearly illustrated by imagining an isentropic chart (say for $\theta = 310°$) where a moist tongue has been entered between two stations; in the axis of this moist tongue a mixing ratio of 7 g/kg has been indicated. Applying the cross section test we see that if the section is of the type shown in the fig. 5a, the moist tongue is real, while in fig. 5b it would be highly improbable, because here the moisture lines are quite arbitrarily drawn.

If the eddies remained stationary and developed in a regular fashion, it would be simple to follow the moist and dry tongues around them. Their life history, however, varies from eddy to eddy, and is closely allied with the energy of the current or currents originally responsible for their development. Once the source of energy of the mother current diminishes, the convergence necessary to maintain the eddy fails and the circu-

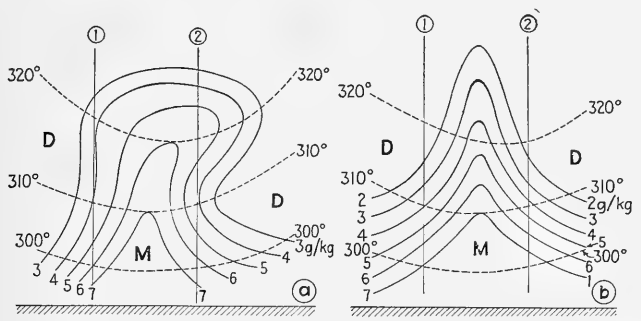

Fig. 5. Illustrating the Cross Section Test for Moist Tongues. (a) Moist Tongue Probable; (b) Moist Tongue High Improbable and Solution to be Abandoned in Favor of Another More Logical One.

lation weakens, finally dissipating or merging into some other circulation. Thus the stability of any particular eddy may, to some extent, be deduced from the distribution of velocity around its center, as shown by Namias [14]. Stable eddies will have well developed circulations with wind velocities increasing radially outward from the center in a nearly symmetrical fashion. Such eddies will tend to rotate as solids. Once the mother current weakens, the symmetry of the velocity profile tends to vanish and the eddy gradually dissipates. Also, for this reason the moisture lines may sometime indicate an eddy pattern which is the result of some already decayed circulation.

By assuming that the smaller anticyclonic eddies are frictionally driven, it is possible to deduce from the distribution of velocity around their centers the general direction of migration. Suppose we have an eddy with a velocity distribution as indicated in fig. 6. This eddy obviously derives its energy from the westerly and northwesterly current, and it may be assumed that the air motion, being nongradient, has a small component in a direction normal to the isobars. In the southeastern quadrant of the eddy there is a region of sub-gradient winds which thus appears as a region of wind divergence. The converging air in the other quadrants (supergradient winds) piling up into the center of the eddy will naturally follow into the divergent region, and the eddy will follow a path indicated in fig. 6. Applying this rule in the general case, we may say that frictionally driven anticyclonic eddies tend to move into the region in which the tangential winds about them are lightest. This direction is normally in the direction of the mother current. It should be emphasized, however, that this rule applies chiefly in the developing stages of the eddy and when it is characterized by a symmetry in the velocity distribution. The use of the above rules for determining the movement and stability of anticyclonic eddies on the isentropic chart will do much to assist in the analysis of the flow patterns and in making forecasts of the movement of moist and dry tongues.

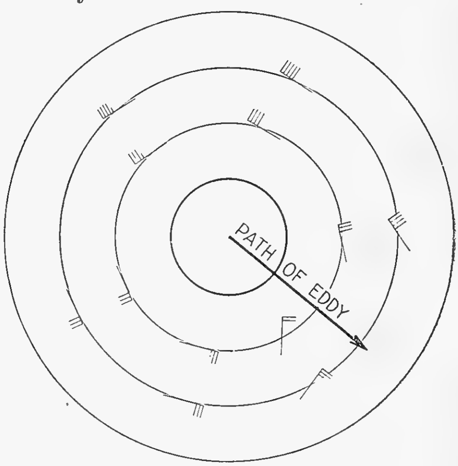

Fig. 6. Relation of Direction of Movement of Anticyclonic Eddies to Distribution of Wind Velocity.

To the left of fast-moving streams a zone of cyclonic wind shear exists, which tends to develop cyclonic eddies. When cyclonic eddies are sharply defined they are generally associated with occluding or occluded cyclones. The dominating current in such cases is not the warm moist air coming from the south, but rather the cold dry air streaming into the eddy cyclonically from the north. The structure of the cyclonic flow pattern normally observed in occluding cyclones is shown diagrammatically in fig. 7. cyclonic eddy observed in tropical currents moving northward. Where the polar air intrudes into the system from the north it develops cyclonic vorticity in order to counterbalance the decreasing cyclonic vorticity of the earth's rotation. Each current attempts to impart the vorticity to its surroundings, and this is accomplished through isentropic shearing stresses. Thus a branch of the moist flow is diverted from the mother cur-

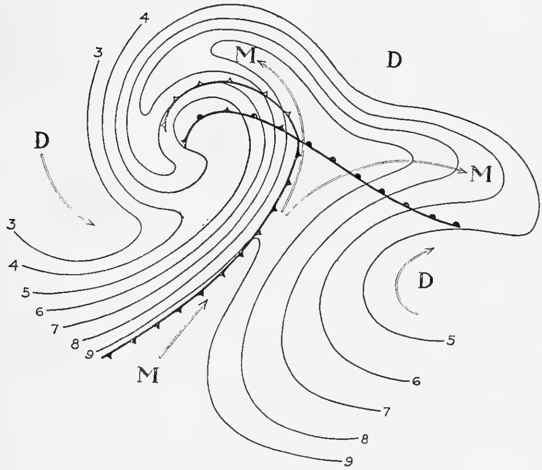

FIG. 7. FRONTS INDICATED AT THE SURFACE AND SCHEMATIC FLOW PATTERN AROUND AN OCCLUDED CYCLONE AS SHOWN BY THE MOISTURE LINES IN AN ISENTROPIC SURFACE IN MID-AIR.

The flow pattern is indicated by the moisture lines, and the arrows represent the instantaneous flow of dry (D) and moist (M) currents relative to the movement of the cyclone. In the model it is observed that two systems are struggling for supremacy: an anticyclonic moist current, M, to the right, and a cyclonic dry current, D, to the left. The moist current, having come up from the south, tends to acquire anticyclonic curvature indicated by the directional flow arrow to the upper right. This part of the pattern may thus be considered as the normal type of anti-

rent into the cyclonic flow. Thus at some point there is branching of the moist flow, and this point appears to be situated in the vicinity of the occlusion point on the surface weather map.

If this branching is due chiefly to lateral shearing stresses, we may arrive at some valuable rules by assuming that in the region of branching *real horizontal divergence* as well as divergence of the stream lines is occurring (Namias [15]). Thus, if the flow pattern prevails through a fairly deep layer of the atmosphere, the surface pressure falls in the region below the branching. This effect is, of course, superimposed upon the pressure changes due to density advection. If the region of divergence is situated some distance from the center of the surface cyclone, a secondary cyclone may form near the peak of the warm sector at the ground.

The effects of the branching upon the pressure distribution at the surface may be roughly estimated by the strength of the interacting currents. Thus a weak anticyclonic eddy will generally yield to an invading strong cyclonic flow of polar air, and in this case no secondary will result, while two strong currents of different vorticity will invariably cause large pressure falls and lead to deepening and possibly cyclogenesis.

§ 5. The Displacement of Flow Patterns with Height

Thus far we have concerned ourselves with flow patterns observed in one isentropic surface. Experience has shown (Simmers [16]) that there is a relatively small difference in the flow pattern from one isentropic sheet to another. This, however, does not hold true if we choose an isentropic surface which is so low that it comes under the influence of the surface friction. The slight displacement of flow pattern with elevation, which occurs above the friction layer, conforms usually with the displacement with elevation of cyclonic and anticyclonic centers. It should be noted, however, that there are exceptions and that these are frequently associated with radical changes in weather situation (Namias [17]). The most important exception, perhaps, is the case when an anticyclonic eddy is present below a cyclonic eddy. This case is represented schematically in fig. 8, where the solid lines represent the flow pattern at some isentropic surface, while the broken lines show the flow pattern along an isentropic surface ten degrees higher in potential temperature. The domes of the isentropic surfaces are indicated by H and the troughs by L. With such a vertical distribution of flow patterns it is clear that while the lower layers over a region are becoming progressively warmer and moister, the higher layers are becoming colder and drier. The advection thus leads to two processes in which the potential energy of the air column is increased: (a) where the lower layers are becoming warmer and the upper layers are becoming colder, the lapse rate is made steeper; (b) since the

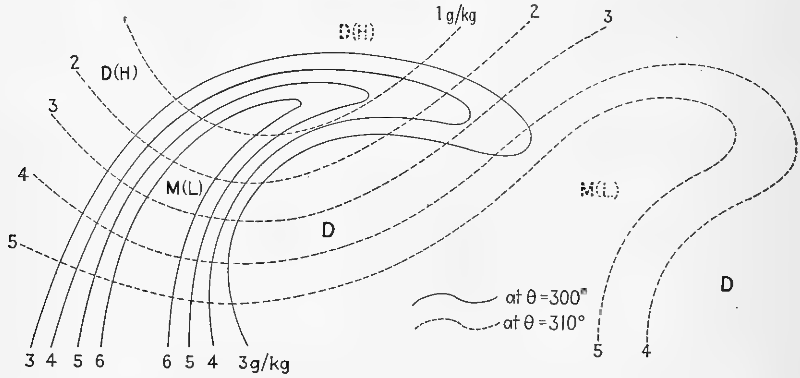

Fig. 8. Flow Patterns at Different Isentropic Surfaces Which Lead to Increasing Instability.

lower layers are becoming richer in moisture while the upper layers are becoming drier, the energy due to convective instability is increasing. This combination of effects makes it easier for frontal activity to produce precipitation, and in general adds to the supply of energy available for cyclogenesis. It also facilitates the outbreak of local showers due to diurnal heating if the moist layer is thick.

A quick test for the conservatism of any particular isentropic flow pattern with height is afforded by the cross sections. Thus in fig. 9A we have the conservative case where it makes little difference in the flow pattern whatever isentropic surface is chosen for analysis, while fig. 9B shows the case in which the flow pattern at a surface $\theta = 305°$ would differ appreciably from that on the surface $\theta = 295°$. In the latter case it is necessary to construct isentropic charts for both these surfaces to obtain a more complete picture of the atmospheric flow patterns.

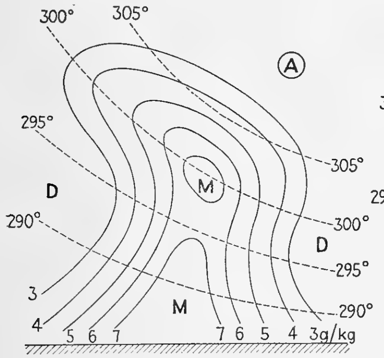

FIG. 9A. A conservative cross-section indicating that choice of isentropic surface will not materially affect location of major flow patterns.

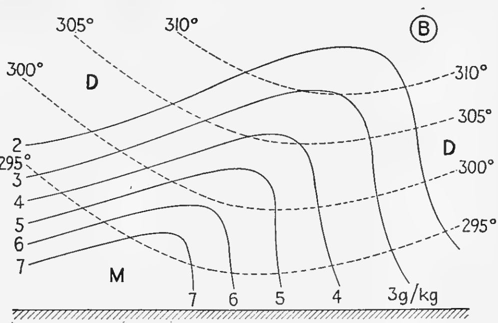

FIG. 9B.—Cross section indicating that flow patterns change appreciably from one isentropic surface to another.

§ 6. The Representation of Gradient Flow in Isentropic Surfaces

The principal reason for the use of isentropic charts is that they afford a means of determining atmospheric motion independent of assumptions regarding the existence of gradient flow. Nevertheless it is frequently desirable to know the gradient flow, and it appears that the mutual adjustment of velocity and pressure distribution takes place in such a fashion that most of the time cross-isobar components of the wind are small compared with the gradient flow.

A highly satisfactory method of representing gradient flow in isentropic surfaces has been suggested by Montgomery [18]. His development leads to the expression for the stream function of the gradient wind in an isentropic surface:

$$\psi = C_p T + \Phi$$

where C_p is the specific heat of air at constant pressure, T is the absolute air temperature at the isentropic surface, and Φ the geopotential. Using meter-ton-second units Φ is expressed in dynamic decimeters, and the value of the constant C_p is approximately 1000.

If lines for equal values of his function ψ are drawn on an isentropic chart, they form streamlines and serve a purpose similar to isobars on a constant level chart.* In regions where pilot balloon wind data are lacking these isentropic stream lines are quite helpful. Moreover, by comparing the patterns of stream lines from day to day, and noting the changes of the ψ values, it is possible to get a better idea as to the future trajectory of dry and moist tongues of the isentropic chart.† There is also considerable advantage in constructing the stream function chart *before* completing the isentropic flow pattern.

*Values of ψ are now being transmitted daily over the airway and Weather Bureau teletype circuits in the United States.

§ 7. The Relation of Isentropic Flow to Precipitation

Since one of the necessary conditions for the formation of precipitation is the presence of sufficient moisture content, it is not surprising that there is generally found some relation between moist tongues and precipitation areas. In winter, when the stratification is relatively stable over the continents, most of the precipitation over continental areas is caused by frontal action. The isentropic chart frequently indicates regions of ascent or descent of air through the relative configurations of the moisture lines.† A frequent type of flow pattern is shown in fig. 10, where a moist tongue ascends the isentropic surface.

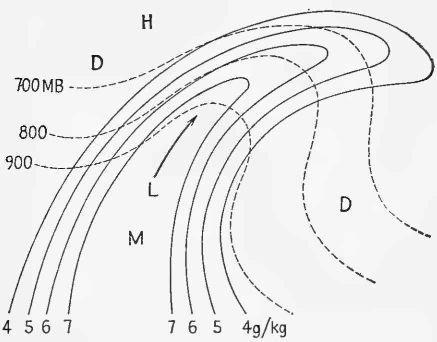

Fig. 10.—Illustrating probable upslope motion of a moist current as deduced from the relation of moisture lines to contour lines.

It should be mentioned that the configuration of moisture and contour lines shown in fig. 10 does not always indicate upslope motion, because the shape of these lines is the result of a lengthy development, and the upslope motion may have ceased by the time when the synoptic picture was obtained.

Another indication of upslope motion along the isentropic surface is obtained from the wind observations. If there are sizeable wind components normal to the contour lines, and if the contour patterns are uniquely defined by observations from a dense network of soundings, then it is *probable* that the air is ascending or descending the isentropic slope in the direction of the wind. However, when the contour lines themselves are displaced with the same speed as the wind component normal to them, the wind components normal to the contour lines are not indicative of up- or downslope motion. If there is any doubt as to whether there is upslope or downslope motion the observed mixing ratios should be compared with the saturation mixing ratios (or the pressure should be compared with the condensation pressure) in order to find out how much lifting is necessary in order to make the air saturated. In addition, consecutive maps should be compared in order to determine whether the air is approaching saturation or not. A study of the closed systems and areas of precipitation will give additional information to this end.

In regions where the heaviest fron-

†See in this connection the interesting suggestions of Starr to show such changes by means of a "relative-motion isentropic chart". (V. Starr, *Bulletin Amer. Met. Soc.*, June, 1940).

tal precipitation occurs the gradient of contour lines is steep and a source of moist air is not far removed. The precipitation band is normally orientated to the left of moist tongues so that there is probably considerable upslope motion of the moist air in this region.

In the warmer seasons much of the precipitation in continental areas is non-frontal in character. This precipitation is chiefly of the convective type, and occurs as local showers and and thundershowers. In Articles VIII and IX we discussed the detailed use of energy diagrams for forecasting these showers. However, the use of energy diagrams becomes even more effective when one takes into consideration not only the state represented by the energy diagram at the time of the sounding but also the probable changes with time caused by the advection of moist and dry tongues at various levels. The isentropic analysis offers by far the most satisfactory method of doing this.

We shall first discuss a few of the characteristic vertical distributions of temperature and moisture normally observed over continental United States in summer. The first type (shown in fig. 11) has an extensive dry layer overlying a relatively moist stratum of about 2 km thickness. The transition zone between the lower moist and the overlying dry air is normally a very stable layer, often a marked temperature inversion. The second type has no discontinuities in temperature and moisture content. Furthermore, the air column is not far from saturation. Type 3 represents a transition between the types 1 and 2, and here there is a 2-km layer of moist air next to the surface, with dry air sandwiched in between this stratum and another layer of high moisture content aloft. As in type 1 there is a stable layer between the dry and moist air, although less stable, and as in type 2 the lapse rate aloft is fairly uniform and normally slightly steeper than the saturated adiabat, while in type 1 it is almost equal to the dry adiabat.
cient to overcome the negative area below and positive areas above for convective impulses from below. Normally such impulses (even at the time of maximum temperature) are insufficient to frustrate the negative area and cause overturning of the whole air column. Type 2 normally gives large amounts of available energy for upward impulses that occur at the

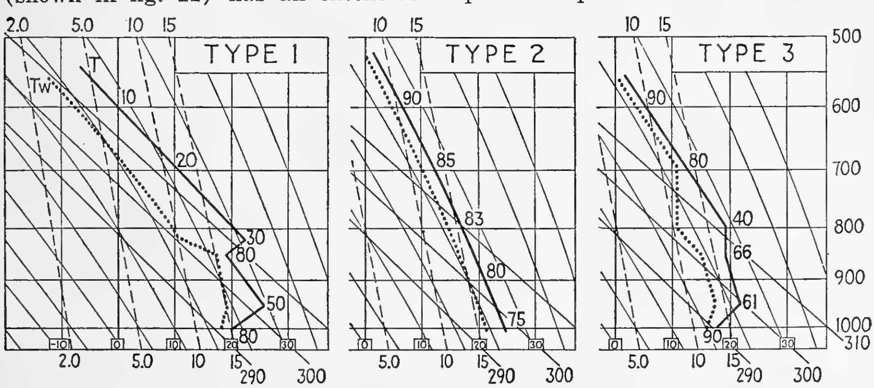

FIG. 11. CHARACTERISTIC VERTICAL DISTRIBUTIONS OF TEMPERATURE AND MOISTURE OBSERVED OVER CONTINENTAL UNITED STATES IN SUMMER. NUMERALS PLOTTED ALONG THE ASCENT CURVES INDICATE RELATIVE HUMIDITY. (Plotted on pseudo-adiabatic charts, $p^{0.288}$ vs T, with the dry adiabats (isentropes) of $\theta = 290°, 300°$A, only.)

time of maximum temperature. This type represents conditional instability of a sort which may easily become realized during the warmer part of the day. Type 3 is very stable for convective impulses near the surface, and normally only negative areas will be observed.

From the above remarks it might be supposed that showers and convective thunderstorms rarely occur with types 1 and 3, while they are common with type 2. While this simple rule would have considerable success in its application to forecasting, it would fail in certain cases. Before we discuss these cases, it is possible to comment further on such vertical distributions of temperature and moisture as are represented in fig. 11. In summer the normal temperature distribution over a large part of the United States is almost barotropic, and the main concentration of solenoids appears to be found along the northern border of the continent*. Consequently, strong westerlies are observed over this portion, and we may look upon the zonal distribution of velocity as being similar to that pictured in the lower half of fig. 2. South of such a westerly current anticyclonic eddies form which create distinct patterns of moisture. This phenomenon occurs so frequently over certain areas of the United States that mean isentropic charts constructed for a month, season, or group of the same seasons of different years, reveal the eddies through the moisture lines (*cf.* fig. 14, e.g.). The normal flow pattern shows that dry air from the north curls anticyclonically southward while moist air from the south converges in a spiral fashion with this dry air into the anticyclonic eddy. The axis of the dry tongue normally runs through the Mississippi Valley, while the moist tongue generally makes its appearance over northern Mexico and then curves eastward.

*See para. 9 *infra* for a more detailed picture of the normal state.

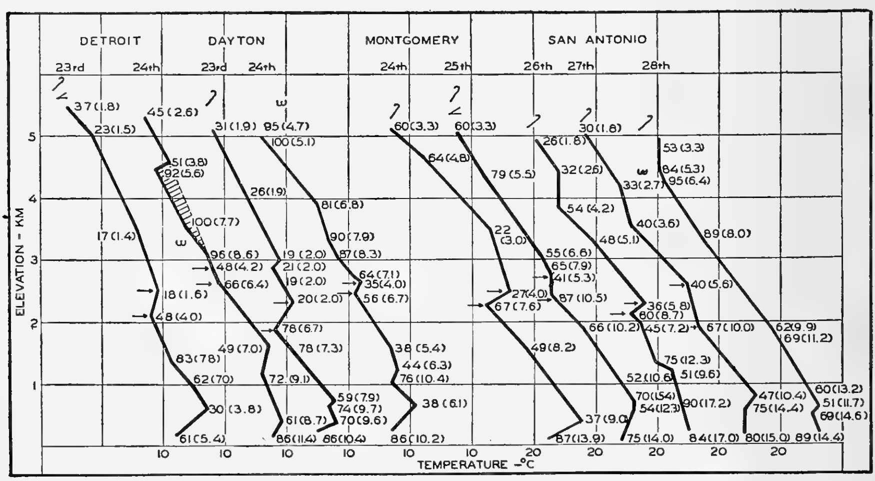

Fig. 12. Destruction of Dry-Type Stable Zones (shown by arrows) After the Advection of a Moist Tongue Aloft. These soundings were made from June 23rd to 28th, 1937. Numbers to the right of the soundings are the relative and (in parentheses) specific humidities. Clouds are indicated by international symbols.

Comparing this pattern with the normal pressure field observed at the surface it will be seen that while the flow of air in the surface layers over eastern United States is uniformly from the south and southwest, there are regions where the current system is reversed aloft. Thus, although the lowest layers of air are generally characterized by considerable homogeneity, there are sections where this moist and warm air, normally of Tropical Maritime origin, is overrun by dryer air coming from the north. Moving southward, the dry air subsides, so that by the time it reaches the core of the anticyclonic eddy it has become warmer and drier than air anywhere in its vicinity. This type of air mass (Ts, or S) is discussed in Prof. Willett's chapter on air mass properties and in the appendix thereto by Mr. Showalter.

Soundings of type 1 may generally be explained on the above basis. The stable transition zones between the moist and overlying dry air are maintained in part by continued subsidence. These stable zones exist for long stretches of time during the summer season, and, when they are especially tenacious, periods of drought result. They are frequently destroyed, however, after the advection of a moist air current aloft. Fig. 12 shows some typical examples when dry stable layers (marked by arrows) were destroyed through advection of moist air aloft. Experience shows that soundings of type 1 will be transformed into type 3 through advection of moist tongues aloft. Through turbulent redistribution of heat and moisture, and through radiative exchange of heat, soundings of type 3 may be transformed into type 2.

Summer showers and thundershowers occur frequently with type 3, while they are rarely observed with type 1. Part of the reason for this is that the stable layer of type 1 effectively damps out any impulses from below, while the same impulses have a better chance of penetrating the less stable transition zone of type 3. But if we consider isentropic mixing and Parr's principle (that this kind of mixing is more pronounced the greater the stability), it becomes clear that in type 1 ascending currents of moist air from below are quickly robbed of their moisture as they enter the stable zone, while in type 3 this effect is not so pronounced. In the case of type 2 the lateral mixing is less pronounced than in 1 or 3 and, moreover, the mixing in this case does not deplete the moisture of the rising current appreciably. Therefore, the condensation levels of type 1 are raised to high levels, and latent heat of condensation is not made available for the growth of Cu clouds and showers. Moreover, the increased lateral mixing with type 1 reduces the total upward momentum of impulses. Cumulus growth is, however, more likely in type 3 and most likely in type 2. The few thunderstorms observed in connection with soundings of type 1 are likely to be high-level thundershowers, and the precipitation from them rarely reaches the ground in appreciable amounts, since it is evaporated into dry air below the cloud base.

In connection with type 3 it has been pointed out that frequently the lapse rate is so stable in the lower layers, even at the time of maximum temperature, that upward impulses are soon damped out. Convective thundershowers which occur with such stratifications normally occur at night-time, especially in the early morning hours. The thunderstorms observed over midwestern United States in summer appear to be mostly of this type. for the diurnal frequency distribution of thunderstorms here shows a night-time maximum at this season. The more probable conclusion is that the impulses generating these storms originate not in the lower layers where the stability is so marked, but in the upper layers. One factor which readily suggests itself as an important one is radiational cooling of the moist air aloft. At the top of these moist currents is normally found another stable zone, frequently an inversion. Cooling of moist air aloft becomes intensified when clouds form at the boundary surface, for clouds act as black bodies. The process thus envisioned demands that convective stirring occurs in the layer cooled from above, just as it must occur in a layer heated from below. However, the processes of removing heat aloft and supplying it from below probably lead to convection of a somewhat different nature. Heating from below, when associated with steep lapse rates, may result in rapid upward motions of small elements, while cooling from above is probably slower acting and is transferred slowly downward.

For a thunderstorm to become energetic and produce sizable amounts of precipitation it must have available an ample supply of moisture. Since in summer the maximum concentration of moisture is generally in the surface layers, it follows that whatever the origin of the convection, overturning must eventually take place throughout these lowest layers if the storm is to become of appreciable intensity. Thus if convection aloft is caused, let us say, by radiational cooling of a cloud layer, we must draw upon some supply of energy to carry this convection into the moisture-rich surface layers. This energy sometimes appears in layers which are conditionally unstable and moist.

The problem of forecasting summer showers therefore depends not only upon the stratification existing aloft in the early morning when soundings are made, nor entirely upon the changes brought about by diurnal heating (see Arts. II, VIII), but also upon the advection of moist and dry tongues aloft. The best method of estimating these advective changes lies in the isentropic analysis. In this manner we are able to detect in advance the likelihood of soundings of any of the above types being transformed into other types through advection.

When reliable isentropic charts are at hand, together with the corresponding cross sections, it is possible to obtain information of forecasting value which is not easily obtained from other charts. The problem of shower and thunderstorm forecasting, therefore, revolves chiefly about the determination of the lapse rate and the source and availability of moisture. Tongues of dry and moist air, as shown by the isentropic charts, may be identified from day to day by means of these charts. In summer it is found almost invariably that thunderstorm activity and showers are associated with the moist tongues, while the dry tongues are free of convective precipitation. Furthermore, by making use of cross sections one is able to form an idea of the representativeness of the chosen isentropic chart. Sources of moisture indicated within the frictional layer may be found to be shallow and not representative of conditions in higher isentropic surfaces. The presence or invasion of dry air aloft would then counteract the possibility of showers and thunderstorms. Thus, though an energy diagram may offer indications of possibilities for shower activity, one should use the isentropic chart to take into account any likely changes. Suppose, for example, that an energy diagram indicates large positive areas and that moist air extends to high levels—both indications of thunderstorm activity during the day. If a tongue of dry air is displacing the moist air at upper levels, the probability of thunderstorms is greatly lessened. Then again, the horizontal extent of the source of moisture must be considered. A narrow jet of moist air will suffer lateral mixing with the dry air flanking it on both sides, and this dessicating process will act against thunderstorm formation. The showers, if they occur at all, will then be restricted to the very central portion of the moist tongue (along its horizontal axis), where the moisture is least affected by the admixture of dry air. On the other hand, extensive regions or broad tongues of moisture may remain comparatively unaltered by lateral mixing, and thereby provide ideal conditions for continued thunderstorm activity.

Thunderstorms caused, at least in part, by radiation from upper levels are best forecast by the use of energy diagrams in conjunction with the isentropic chart. The first step is to decide what changes in the temperature and moisture distribution are likely to take place. The changes in the flow pattern of moisture are brought about mainly through advection and lateral mixing. Wind directions and velocities on the isentropic chart provide the chief indices of the magnitude of both these factors. For example, lateral mixing is most favored in regions where the horizontal wind shear is greatest. After considering modifying factors, it is necessary to determine the layer from which the principal radiational loss of energy will take place. This layer is not difficult to place; it is usually the most pronounced dry inversion of the sounding. If it appears at low levels, in general below about 2 km, it will be of little significance in helping large-scale convective activity, as was explained in the discussion of the lateral mixing in the dry-inversion. But the radiational emission layer, when at higher levels, becomes increasingly important, for the heat lost to space at the boundary helps set off convection through a thick layer of atmosphere. The rate of cooling at cloud tops is appreciably greater than from unsaturated air under the same conditions, and therefore the nearness to saturation of the moist layer must be considered.

If the lower layers of the atmosphere are too cold, the tephigram will indicate that convective energy aloft will be dissipated before it can receive supplies of moisture from the lower layers. In this case, even though the chief emission layer is at high levels, thunderstorms are not likely to occur.

§ 8. The Processes which Tend to Disrupt the Continuity of Isentropic Analysis

From the standpoint of following the same sheet of air from day to day the ideal method of representation would be one in which non-adiabatic as well as adiabatic influences were taken into consideration. With such a method one could construct charts along substantial sheets*—that is, sheets which contain the same air particles from day to day. By following these identical sheets and describing the motion of elements with respect to such sheets, one would be using the Lagrangian method of

*Other writers have used the expression "equi-substantial sheet" instead of "substantial", when referring to a fluid sheet rather than to a surface of a solid; but it does not seem necessary to make this fine and cumbersome distinction.—*Ed.*

describing the changes in the atmosphere. While at present it is not possible to chart exactly substantial surfaces, there is much evidence to indicate that isentropic surfaces do not depart appreciably from substantial surfaces. Thus, the isentropic method of analysis is essentially a Lagrangian method. In a study of subsidence Namias [20] has shown that the potential temperatures at the bases and at the tops of subsidence inversions remain fairly constant from day to day. Thus, if, in such cases, isentropic surfaces are used, it is reasonably certain that we are dealing with the same sheet of air particles from day to day. Moreover, it is frequently observed that sandwiched layers of dry and moist air remain within the same isentropic sheets for several days. From these observations it appears that, as a first approximation, we may consider isentropic surfaces as substantial surfaces.

Nevertheless, there are always at work non-adiabatic processes which tend to destroy the conservatism of isentropic surfaces and to raise or lower the isentropes relative to the

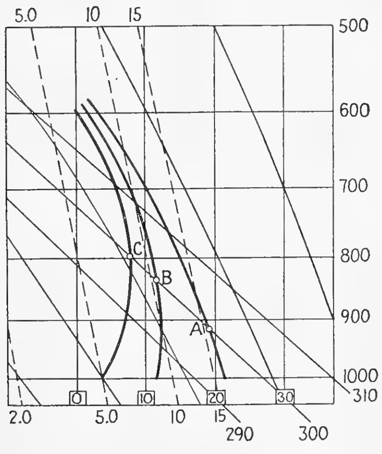

FIG. 13.—Illustrating the influence of non-adiabatic cooling on the elevation of an isentropic surface.

substantial surfaces and also tend to transport moisture across the isentropic surfaces. These non-adiabatic processes are mainly due to: (a) radiation, (b) evaporation and condensation, and (c) convection. While the influences of these processes may be appreciable over lengthy intervals of time, they are usually insufficient for disrupting the fundamental isentropic flow patterns from one day to the next.

The influence of radiative cooling is illustrated in fig. 13. As cooling proceeds, the temperature distribution changes from A to B to C. The substantial surfaces do not change elevation but since the temperature decreases, the height of any given isentropic surface increases from day to day. Since the normal moisture distribution is one in which mixing ratio decreases with elevation, the mixing ratio observed in a given isentropic surface decreases with time. Since the rate of radiational cooling in the free atmosphere is usually small compared with the adiabatic cooling, it does not destroy the essential character of the flow pattern. This slow rate of free-air cooling is indicated by the cooling curves computed by Möller [21], which suggest that the *mean* temperature change resulting from the radiative unbalance in the atmosphere hardly exceeds 1.5C° per day,* which with normal lapse rate corresponds to a vertical displacement

*Elsasser (Unpublished MS) has recomputed such cooling curves on the basis of newer data on the water vapor absorption spectrum. In general his values do not differ excessively from Möller's. However, both Möller's and Elsasser's curves are based on monthly means of upper-air conditions. On individual days the cooling must be much greater sometimes, perhaps as much as 10° or 15° C per day from a saturated warm stratum with a deep dry inversion above it. The figure 1.5° C quoted is for mid-latitudes; Elsasser's (*Bull. Amer. Met. Soc.*, May 1940) *mean* values for *Florida* soundings even in winter show over 2.0° C per day cooling at moderate elevations. Elsasser has published a radiation chart with which radiation in individual soundings can be computed (Calif. Inst. Techn. 1939).—*R. G. Stone.*

of the isentropic surface of about 300 m in one day. Nevertheless, it is at times necessary to introduce this factor to explain changes in moisture content or height of the isentropic surface which cannot satisfactorily be explained by advection or other causes. Where the isentropic surface is not far from snow-covered mountains, cooling by radiation and eddy transfer of heat must frequently be considered.

Opposite effects are observed when there is non-adiabatic heating. Then a substantial surface has its potential temperature raised so that the isentropic surface is lowered, and since the specific humidity normally decreases with elevation, there appears to be an increase in moisture within the affected area of the isentropic chart. This type of non-adiabatic modification is significant when the isentropic surface is near the ground. The influence is greatest in continental areas during the summer season. It is also important in mountainous country during all seasons.

Let us consider next the non-adiabatic effects produced by evaporation and condensation. Since condensation liberates and evaporation consumes heat, it becomes important to know whether these processes occur above, within or below the isentropic sheet under discussion.

If the chosen isentropic surface lies above the region where condensation occurs, its characteristics will not be materially affected. An example is afforded by the instability snow showers of polar continental air masses of winter. These flurries are generally formed in a shallow layer of air next to the earth's surface—a layer which is far below the representative isentropic surfaces which are chosen so as not to intersect the ground even in tropical air. In this case isentropic surfaces remain practically substantial surfaces.

If condensation and precipitation set in within the chosen isentropic sheet, latent heat is liberated and the potential temperature of the substantial surface is raised. The isentropic surface is then found at lower levels, and, since the specific humidity normally increases downward, the specific humidity in the isentropic surface increases. This increase might erroneously be interpreted as being due to advection from a neighboring source of moisture.

When precipitation falls through an isentropic sheet which is not saturated with moisture, evaporation will cool the air while the specific humidity increases. This lowers the potential temperature of the substantial surface, and raises the isentropic surface. If the moisture content decreases with elevation the mixing ratio at the chosen isentropic surface decreases in proportion to the humidity gradient. On the other hand, if the moisture content increases with elevation (as it often does along well defined frontal surfaces) the mixing ratio will increase. The increase in moisture with elevation is usually so slight that, though the isentropic surface rises, little change in the pattern of moisture results.

Probably the most significant process at work in causing isentropic surfaces to depart from substantial surfaces is convection, because vertical currents are highly effective in transporting moisture to higher levels. If it were not for the replenishment of moisture by convective currents, the moist tongues which are not associated with upslide motion along frontal surfaces would soon be dissipated through lateral mixing with the dry air flanking them.

In the discussion of the influence of convection on the moisture patterns on the isentropic chart it is convenient to distinguish between widespread convection within unstable air masses and convection associated with fronts. Convection that occurs along fronts (notably cold fronts) is usually restricted to a narrow zone which coincides with a moist tongue on the isentropic chart. While the vertical currents transport moisture to higher levels, water is also precipitated from the frontal cloud system. The convection which occurs along such fronts merely replenishes the moisture content of the moist tongue, and it does not disrupt the continuity of the isentropic analysis. The same also applies to purely local convection ("pinpoint convection"). The matter may, however, be different in cases of widespread convection. The convective transfer of moisture may then at the beginning of this process radically change the moisture pattern aloft. This is particularly the case in Polar continental air when it moves over an ocean, because the convective currents will then transport much moisture to high levels. The convection that occurs in unstable Polar continental air moving over land in winter does not usually reach up to the representative isentropic surface; it therefore does not appreciably influence the moisture pattern aloft.

From the above it follows that the occurrence of convective as well as frontal precipitation is closely related to the moist tongues, and that the isentropic charts afford the best means for analyzing the processes in the free atmosphere.

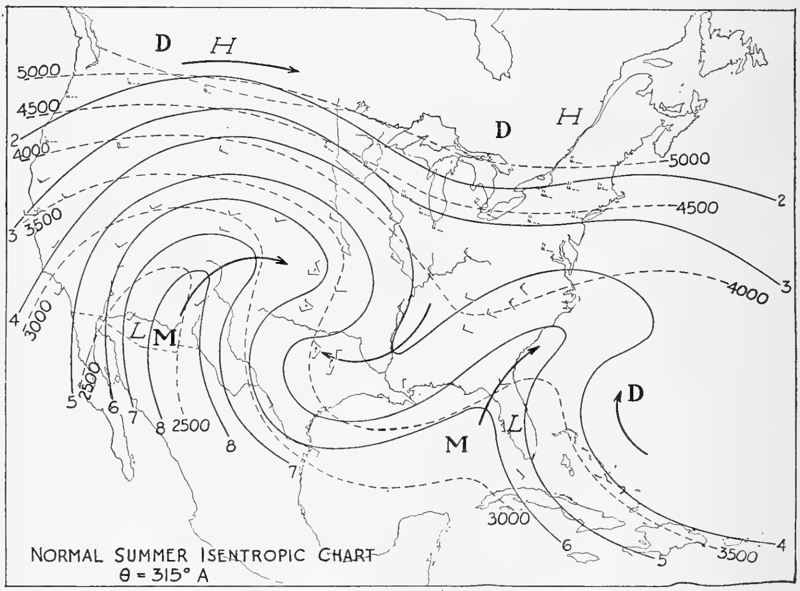

FIG. 14. THE NORMAL FLOW PATTERN OVER UNITED STATES IN SUMMER. This chart is based upon aerological data for summer months from July, 1934, through August, 1939. It was constructed by averaging values interpolated at 5 degree intersections of longitude and latitude from analyzed monthly mean isentropic charts. The winds are normal resultant upper-air winds.

§ 9. The Mean State of the Atmosphere as Revealed by Isentropic Charts

If the daily aerological soundings during a given month are averaged for various stations, we may construct isentropic charts and cross sections representing the mean state of the atmosphere for that month. The mean air flow may be obtained by computing the resultant winds from pilot-balloon observations. (See the mean charts published monthly in the *Mo. Wea Rev.*). Similarly, it is possible to construct mean seasonal isentropic charts, and, if data were available, normal charts. An example is shown in fig. 14.

The outstanding feature of the normal summer pattern is the existence of two well-defined anticyclonic cells — one centered over western Texas, the other somewhere off the southeastern coastal states. Some light on the question of the formation of these eddies is furnished by north-south atmospheric cross sections, a typical one for the summer season being reproduced in fig. 15. This section brings out the well-known fact that over North America the principal summertime Polar front is generally found in the vicinity of the Canadian 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 [22] the eastward current in the homogeneous air is maintained by frictional stresses from the much stronger westerly current to the north, and this energy is continually being dissipated in the form of eddies further to the south. 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 upper-air winds. The mean isentropic charts for individual summer months invariably reveal such eddies, and most of the time there are observed two anticyclonic cells placed about as they are in fig. 14.

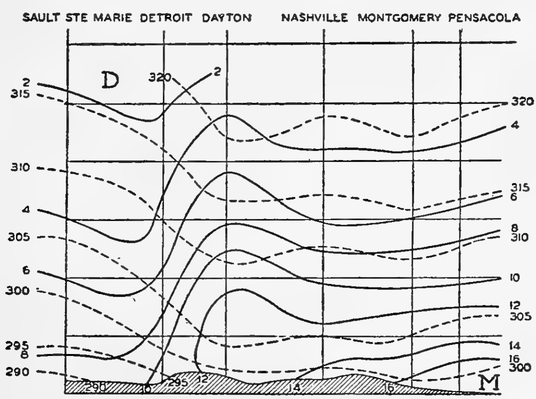

Fig. 15.—North-south vertical cross section for August, 1936, showing the distribution of potential temperature (broken lines) and specific humidity (full lines).

In all the mean summer isentropic charts studied there has been a moist tongue projecting from northern Mexico recurving towards the northeast and east. The region covered by this tongue has a maximum of precipitation in summer. It has been suggested by Wexler [23] that this preferred site for the anticyclonic eddy is largely due to topography and results from the field of solenoids to the north as well as from the field of solenoids established between the warm Rocky Mountain region and the colder Pacific region.

The point of injection of the moist tongue of the eastern eddy is normally over Florida, in which region there is a summer maximum of rainfall and a high frequency of thunderstorms.

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 state are always associated with considerable anomalies in rainfall and temperature (Wexler and Namias [24]). Thus during the month of August, 1936, when one of the most severe droughts and heat waves occurred in the mid-west, the mean monthly pattern showed only one very extensive anticyclonic eddy covering the entire country rather than the normal double cellular pattern.

In winter, when the mid-latitude solenoid field is situated farther to the south, the mean isentropic charts do not generally display any outstanding eddies. Simultaneously, the daily charts show pronounced eddies with moist and dry tongues. The individual eddies then move rapidly, without preferring any particular site. While the above refers to the conditions over the North American continent, there can be little doubt that the principles outlined here apply in a general way throughout the world.

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Isentropic Analysis of a Thunderstorm Situation, June 22-27, 1937

The following series of charts was used by Mr. Namias to illustrate his epoch-making paper on thunderstorm forecasting with the aid of isentropic analysis (*Bull. Amer. Met. Soc.*, Jan. 1938).

The tephigram in Figure 1 failed to give evidence of liability to showers since there was a large negative area, but this was in the warm sector (Fig. 3) and thunderstorms actually occurred in the vicinity that afternoon. Figure 2 gave every indication for showers but none occurred anywhere near the station that day. Therefore the assumption that any one particle or mass of air rises adiabatically through a resting medium and that airplane soundings taken in the early morning are representative of upper air conditions for the rest of the day are clearly invalid in some cases such as these. A study of these cases with the isentropic charts and cross-sections shown in Figures 4-7 permits these exceptions to be foreseen. The isentropic charts (Figs. 4 and 6) show a large moist tongue advancing over Detroit on the 23rd-24th which rapidly increases the specific humidity by many grams (see Fig. 9, also Fig. 12 of Mr. Namias' chapter on Isentropic Analysis) so that showers could occur.

The advection of this moist air at intermediate levels destroys the preexisting dry-type inversions (see Fig. 9, below) and results in less stable lapse rates in these layers. Consequently parcels of air rising from the surface into this region will (by Parr's principle) undergo less lateral (isentropic) mixing with the environment. Moreover, since the convection now takes place in a generally moist stratum, the condensation levels of rising surface particles are low enough for upward pulses to reach, thus making available for convection the heat of condensation. In this way the advance of a moist tongue aloft may be accompanied by a train of showers; this interpretation is given in Fig. 8 for the period under discussion. The forecasted showers at San Antonio on the 27th failed to occur (Cf. Fig. 12, Namias' Article X) because the center of the moist tongue passed somewhat to the north and the Gulf influence kept the lower levels too cool and stable (showers did occur farther west, however).—*R. G. S.*

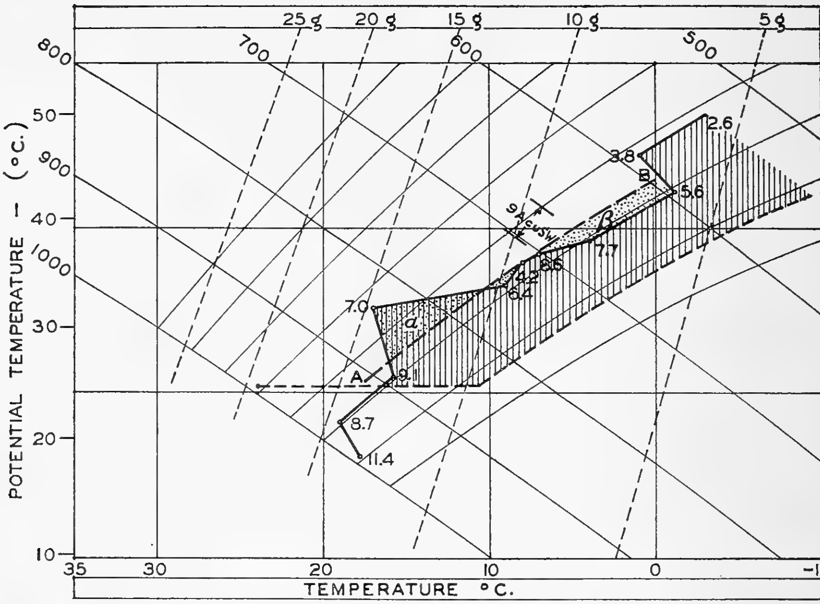


FIG. 1. TEPHIGRAM OF THE AIRPLANE SOUNDING AT DETROIT, MICHIGAN, JUNE 24, 1937, BEGUN AT 4:00 a.m., E. S. T. (Specific humidities entered beside the solid line of the sounding. The dotted line is the path of a particle rising from the surface after it attained the maximum temperature of that day. The broken line AB is a pseudo-adiabat, the significance of which is explained in the text. Areas horizontally shaded represent positive energy while those shaded vertically indicate negative energy, which opposes convection.)

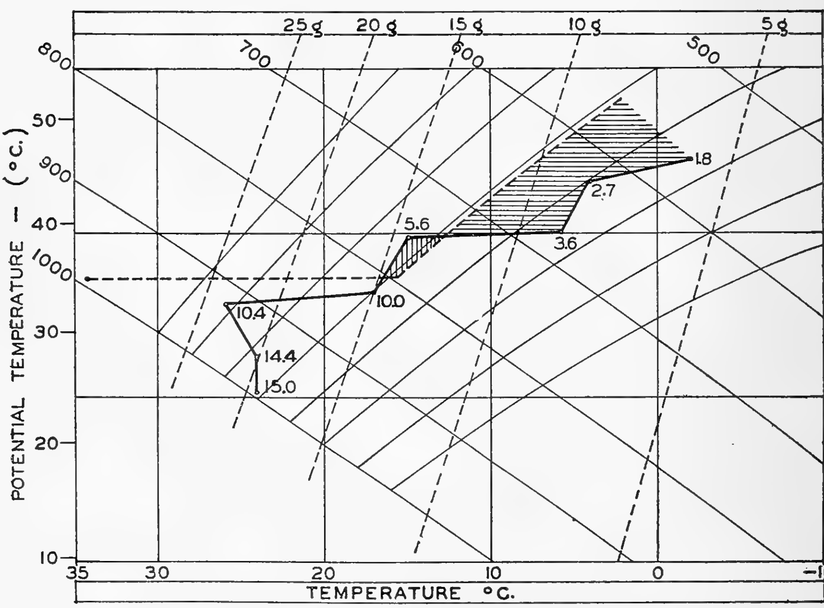


FIG. 2. TEPHIGRAM OF THE AIRPLANE SOUNDING AT SAN ANTONIO, TEXAS, JUNE 27, 1937, 2.00 a.m. E. S. T. (For explanation see legend to Fig. 1.)

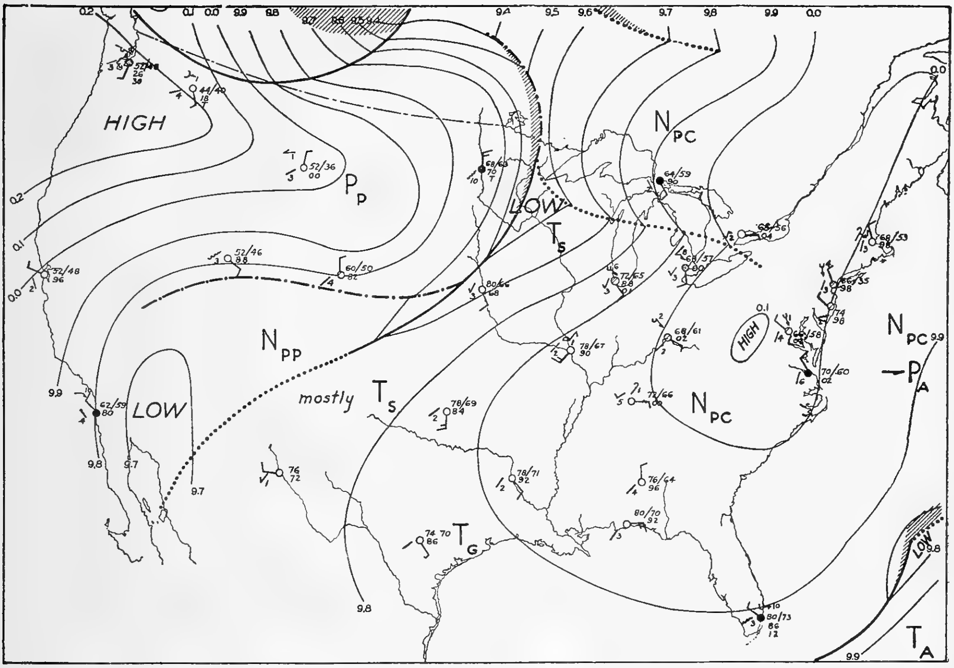


FIG. 3. SYNOPTIC CHART OF SURFACE WEATHER OVER UNITED STATES AT 7:30 a.m., E. S. T., JUNE 24, 1937. (Cold fronts are indicated by heavy solid lines, warm fronts by dotted lines, and occluded fronts by alternately dashed and dotted lines. The hatching indicates areas where precipitation is falling at the time of observation. Air-mass symbols are those customarily in use in the United States and introduced at the Massachusetts Institute of Technology. Surface observations at or near aerological stations [circles] are entered in the customary manner. To the right of the circles from top to bottom: temperature and dew point (°F), pressure, and precipitation. Winds are indicated by arrows, the numbers of half-barbs corresponding to Beaufort numbers of force. To the left of the station are the clouds in international symbols, and the pressure characteristic and change in the preceding three hours.)

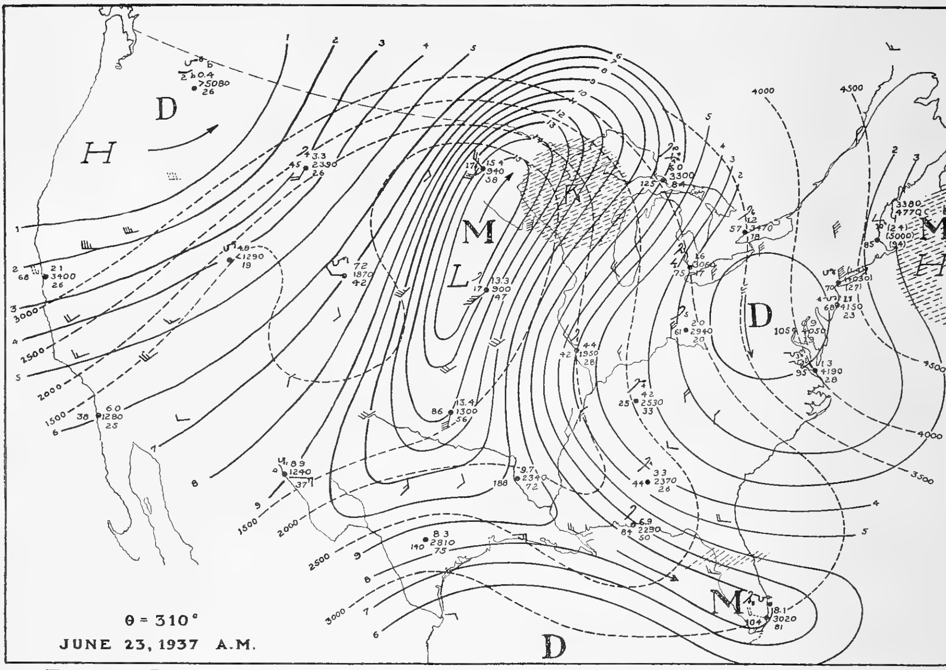


FIG. 4. ISENTROPIC CHART CONSTRUCTED FROM AEROLOGICAL OBSERVATIONS MADE IN THE EARLY MORNING OF JUNE 23, 1937. (The explanation of the construction and the legends is given after Fig. 7.)

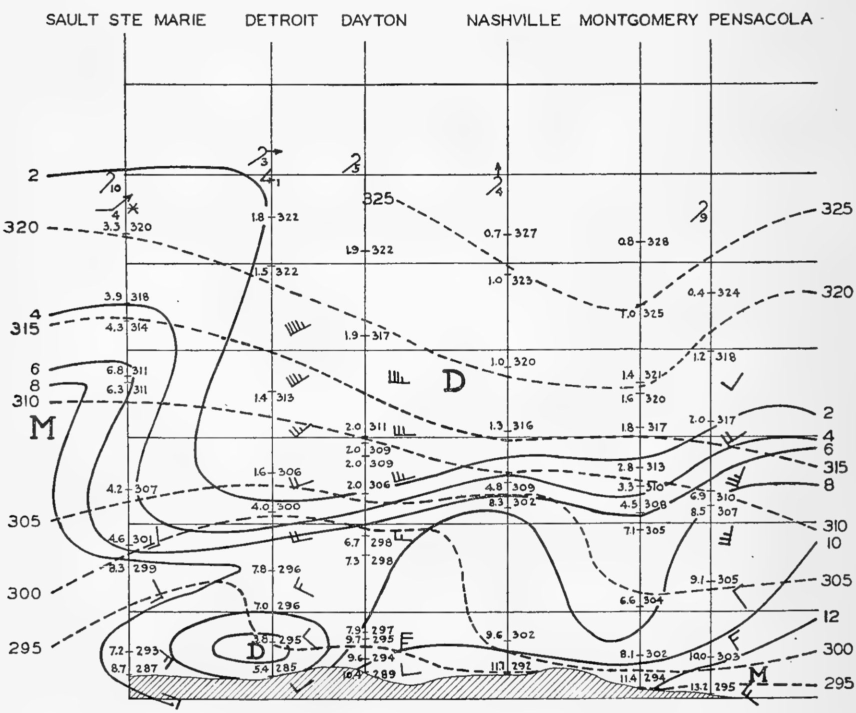


FIG. 5. VERTICAL CROSS-SECTION THROUGH THE ATMOSPHERE FROM ST. STE. MARIE, MICHIGAN, TO PENSACOLA, FLORIDA, ON JUNE 23, 1937, EARLY A.M. (The explanation of the symbols is given after Fig. 7.)

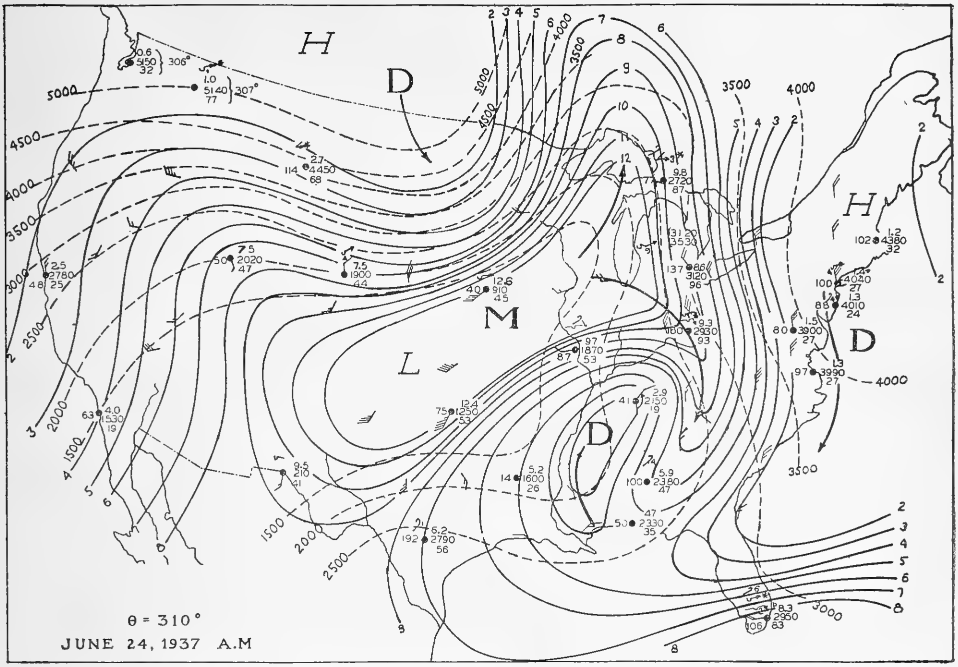


FIG. 6. ISENTROPIC CHART CONSTRUCTED FROM AEROLOGICAL OBSERVATIONS OF THE EARLY MORNING OF JUNE 24, 1937. (For explanation of construction and symbols after Fig. 7.)

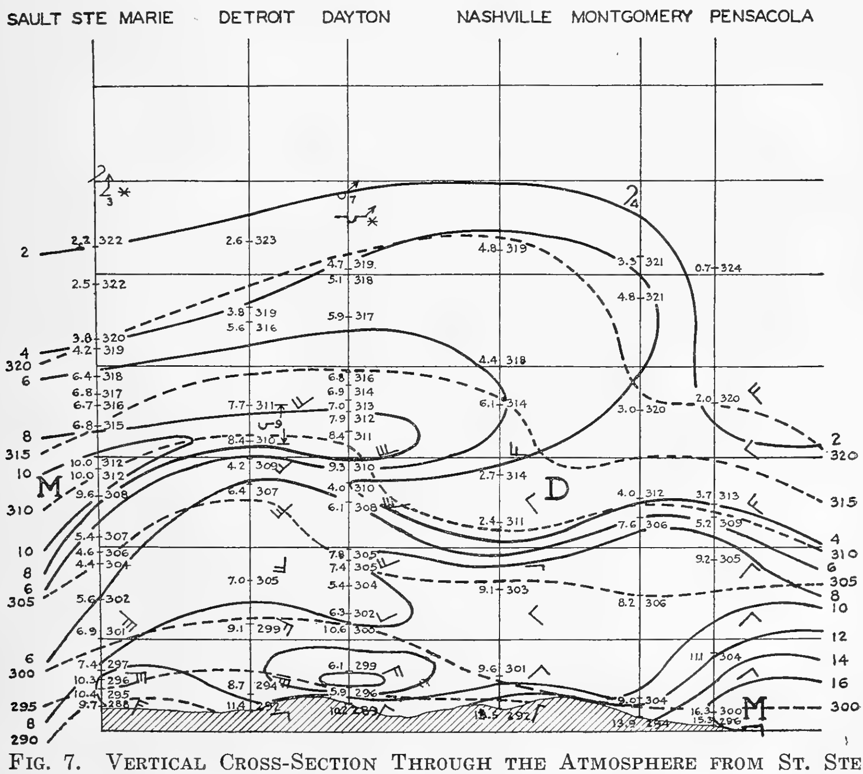


FIG. 7. VERTICAL CROSS-SECTION THROUGH THE ATMOSPHERE FROM ST. STE. MARIE, MICHIGAN, TO PENSACOLA, FLORIDA, ON JUNE 24, 1937. (The explanation of symbols is given below.)

Explanatory Legends For Figures 4, 5, 6 and 7

Figures 4 and 6 are charts constructed upon a surface of constant entropy (constant potential-temperature). For these cases the particular value of potential temperature chosen was 310°. Solid lines are lines of specific humidity while dotted lines are lines of elevation (in meters) of the chosen surface above sea level. "M" signifies the center of a moist tongue, "D" the center of a dry tongue. "H" and "L" are entered at the crests and troughs of the isentropic sheet. The numbers to the right of the aerological stations are, in the order listed, the specific humidity, the elevation, and the relative humidity at the given isentropic sheet. To the left of the station is entered the pressure, in millibars, of the layer bounded by the 305° and 310° surfaces of potential temperature. The change from day to day in these values offers indications of convergence and divergence. Winds at the isentropic surface are shown by arrows, the number of half-barbs being roughly equivalent to Beaufort numbers of the scale of wind force. Clouds are indicated by the international cloud symbols. Subscripts give the amount of cloud and arrows the direction of movement. If the letter "b" appears following the symbol the clouds are below the isentropic sheet, if elevations are given the clouds penetrate the sheet, and if nothing is appended they are above it. An asterisk (*) indicates that the clouds are within the range of the sounding, yet no height was recorded. Heavy arrows represent the probable path of the tongues with respect to the isentropic surface (that is, choosing a coördinate system fixed to the isentropic surface).

Figures 5 and 7 are vertical cross-sections extending north-south from Sault Ste. Marie, Michigan, to Pensacola, Florida. The evenly spaced horizontal lines represent full kilometers of height. Solid lines are drawn for specific humidity, dashed lines for potential temperature. The humidities in the individual soundings are entered to the left and the potential temperatures to the right of the vertical. Winds are drawn so that north is to the left of the cross-section. For example, at 3 km over Dayton on the 23rd (fig. 5) there is a N wind of force six. Clouds are indicated, as in figures 4 and 6, by the international system; those above the soundings are either above the top of the ascent, or, if accompanied by an asterisk, within the levels penetrated by the sounding but not placed specifically. Here again "M" stands for moist, "D" for dry.

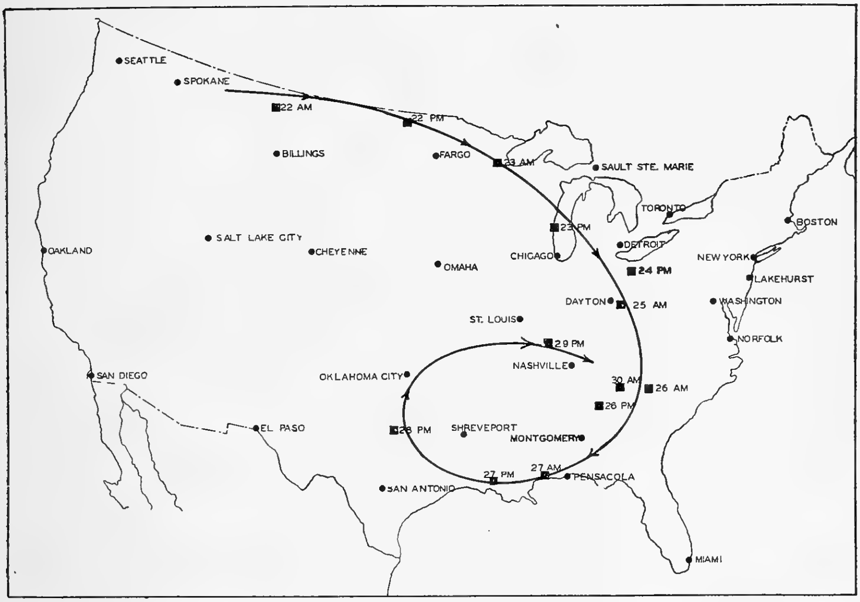


FIG. 8. TRAJECTORY OF THE CENTER OF MAXIMUM THUNDERSTORM ACTIVITY FOLLOWING THE INVASION OF A TONGUE OF MOIST AIR ALOFT; AND POSITIONS OF AIRPLANE SOUNDING STATIONS USED IN THIS ANALYSIS (circles). (The black squares mark the successive positions of the center of thunderstorm activity for the twelve-hour periods preceding the dates entered beside them; these were fixed with the aid of thunderstorm reports and 12-hourly amounts of precipitation.)

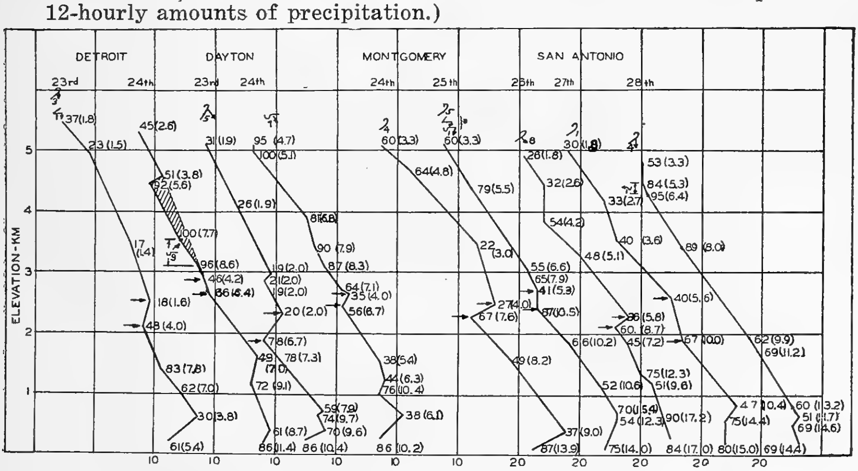


FIG. 9. LAPSE-RATE DIAGRAMS OF SOUNDINGS DISCUSSED IN THIS REPORT. (The scales are such that a 45° line sloping upwards from right to left would represent the dry adiabat. Numbers to the right of these soundings are the relative and (in parentheses) specific humidities. Clouds are indicated by the International symbols of 1932. Arrows mark off dry-type moisture discontinuities.)

Analysis of the Rainfall Situation over the Western States May 6-7, 1938, by Means of Air Mass and Isentropic Charts

The surface maps and isentropic chart below are from the article by Mr. R. H. Weightman in the *Bull. Amer. Met. Soc.*, April 1939, and were analyzed at the U. S. Weather Bureau, Washington. The isentropic chart illustrates the use of condensation pressures practiced by the Bureau. The relation of the surface fronts and of the isentropic flow and moisture distribution to the rainfall patterns can be readily interpreted in light of the principles outlined by Mr. Namias. The rain along the Gulf coast is accredited to prefrontal instability due to convergence ahead of the cold front over Texas; the rain over the Missouri and Mississippi valleys is definitely frontal; the northern Rocky Mountain rain area is due to instability in a moist tongue ascending from the south (in the northwest the 303°-surface was above the tops of the soundings on May 7); the Colorado-New Mexico rain belt is related both to the surface fronts and to the ascent of the moist tongue. The eastward advance of the moist tongue brings rain over Iowa on the a.m. of the 7th.—*R. G. S.*

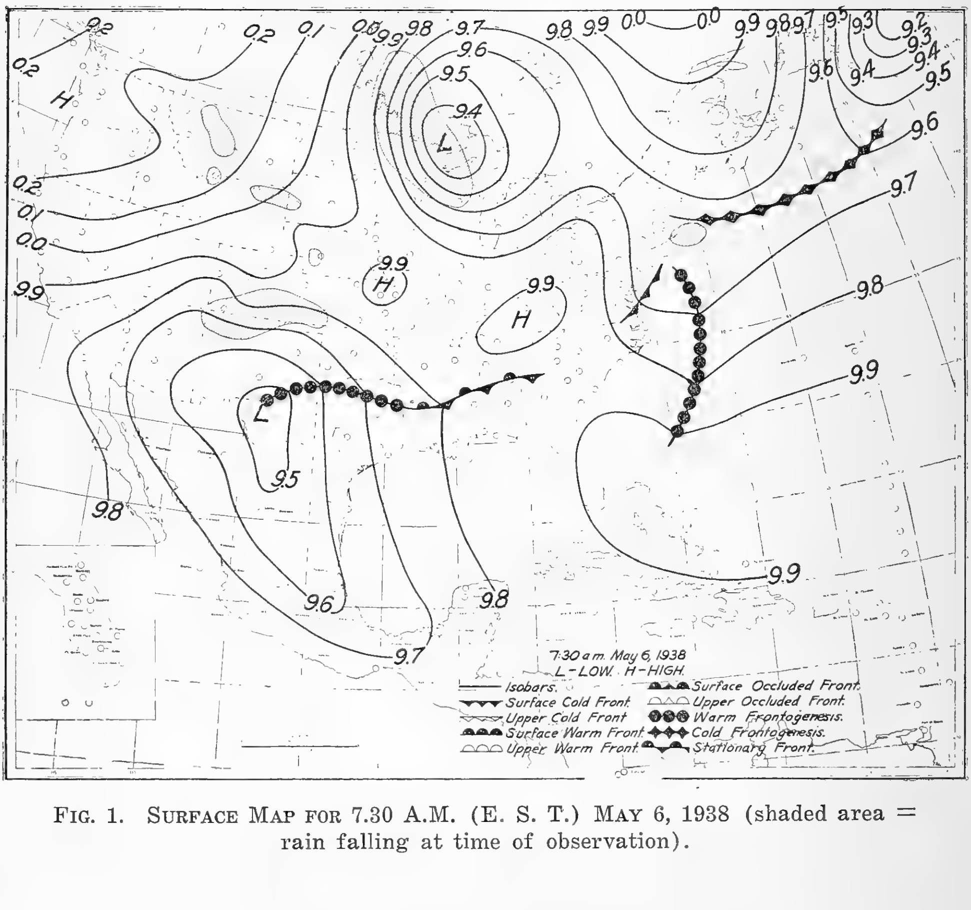


FIG. 1. SURFACE MAP FOR 7.30 A.M. (E. S. T.) MAY 6, 1938 (shaded area = rain falling at time of observation).

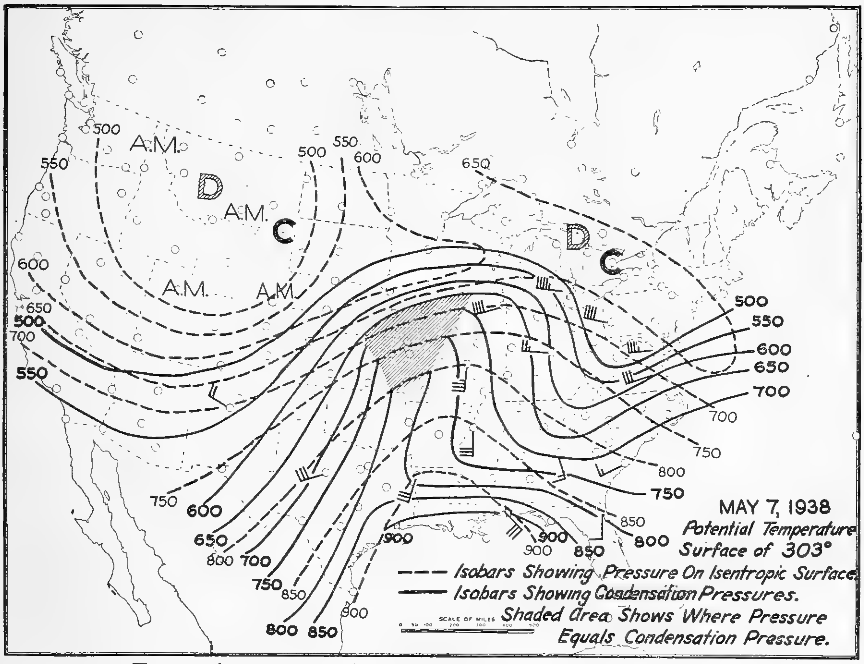


FIG. 2. ISENTROPIC CHART FOR 5.30 A.M., MAY 7, 1938

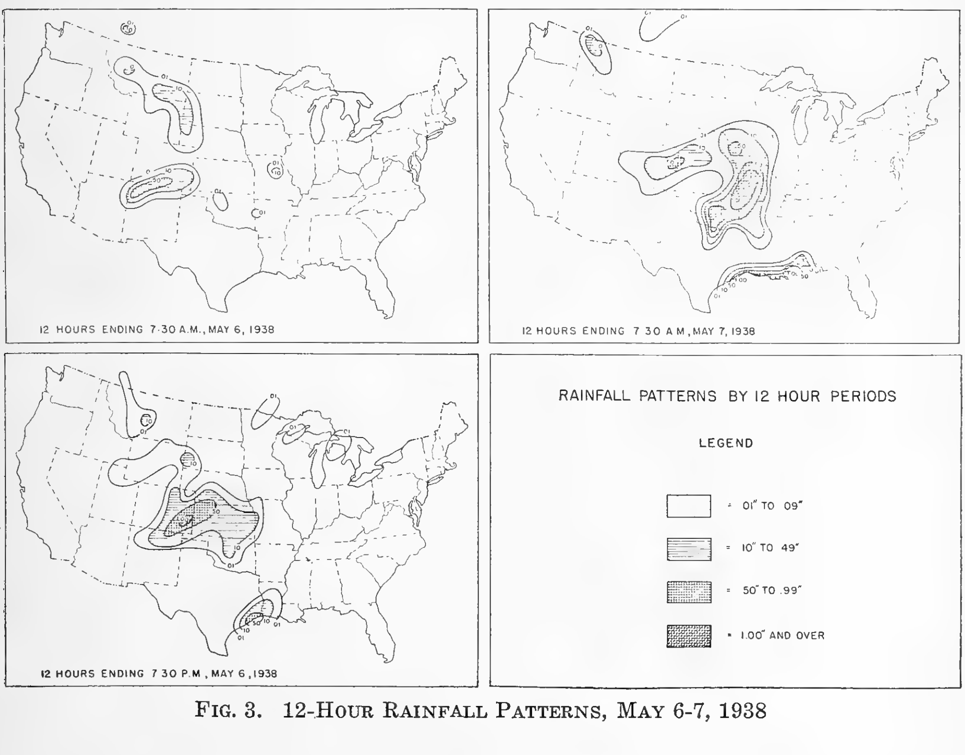


FIG. 3. 12-HOUR RAINFALL PATTERNS, MAY 6-7, 1938

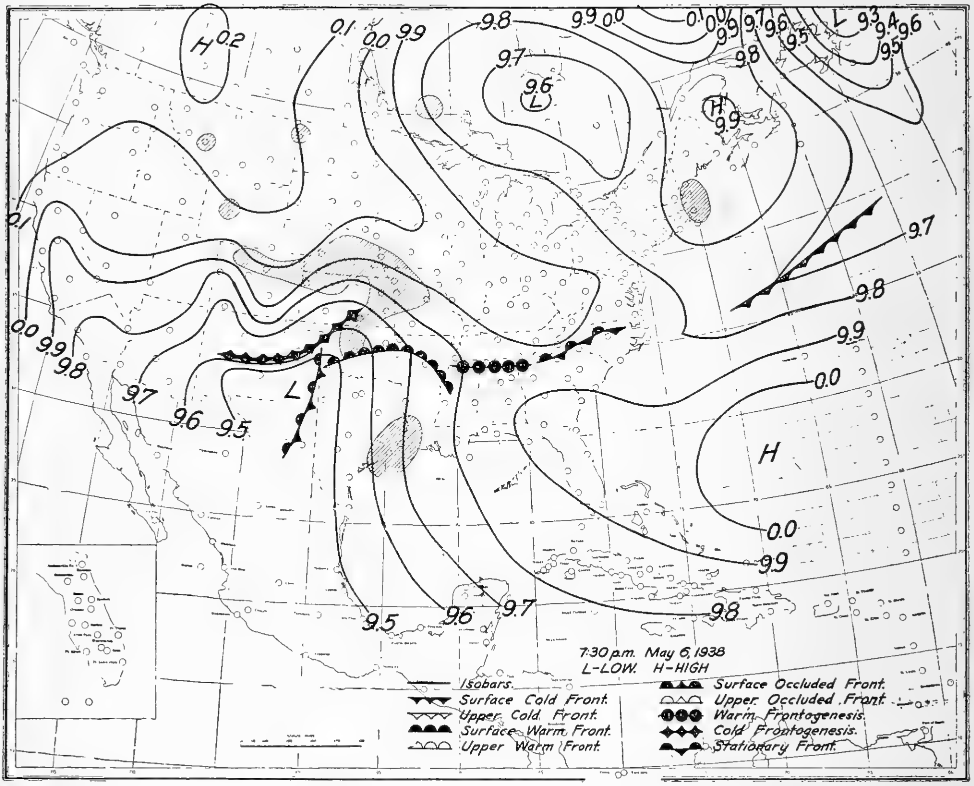


FIG. 4. SURFACE MAP FOR 7.30 P.M. (E. S. T.) MAY 6, 1938.

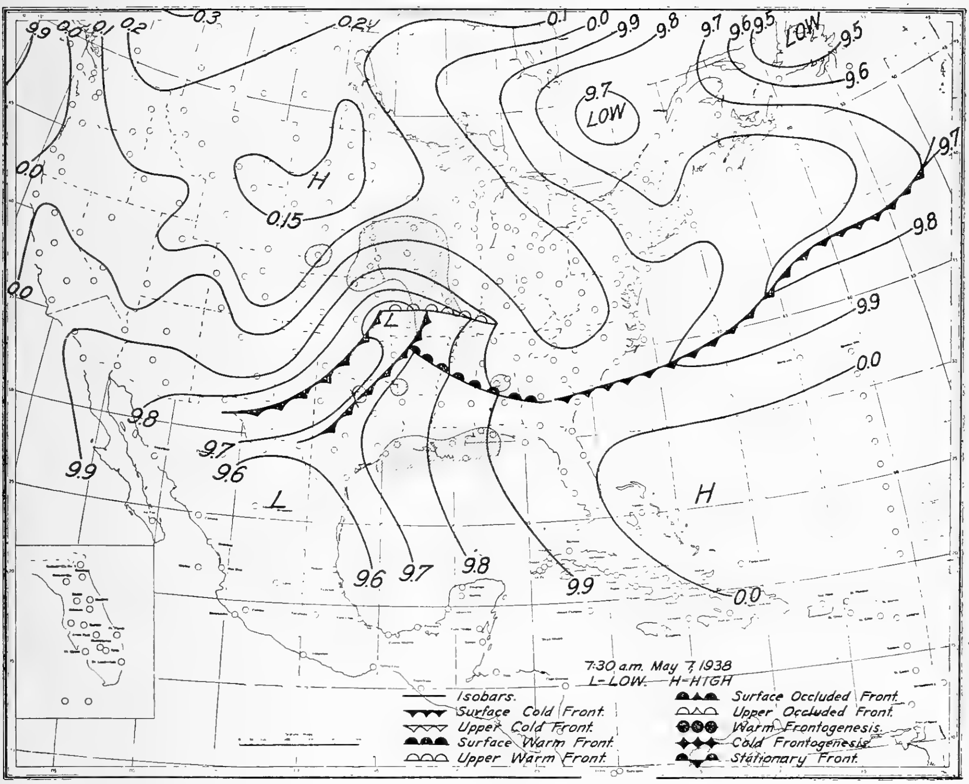


FIG. 5. SURFACE MAP FOR 7.30 A.M. (E. S. T.) MAY 7, 1938

Examples of Upper-Air Cross-Sections Showing Interpretations of the Tropopause, etc.

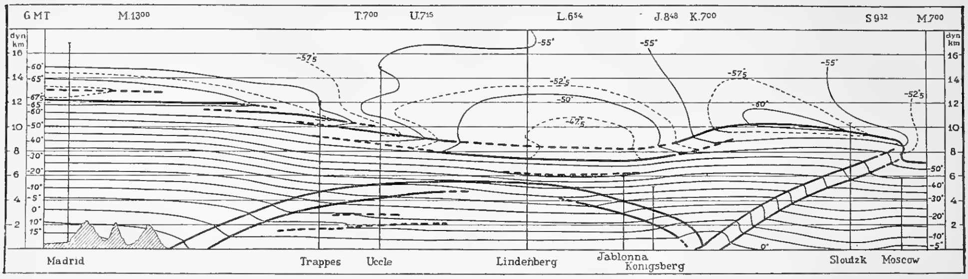


FIG. 1. Vertical cross-section from Madrid to Moscow, A.M. of Feb. 17, 1935. Isotherms in °C, fronts and tropopauses heavy black lines; dashed lines for subsidence inversions (troposphere) and for indefinite tropopauses (being formed or dissolved). The overlapping of the multiple tropopauses is called for by the theory of Palmén, that the adiabatic cooling or heating from the pumping effect of passing disturbances and highs in the troposphere causes the tropopause to be destroyed at one level and reform at higher or lower levels as the case may be. Only a few soundings (vertical ordinates) were available for this analysis by Bjerknes and Palmén (*Geofys. Publ.*, v. 12, no. 2, 1937, p. 52) but experience from other cases analyzed permits the interpolations. The fronts are drawn double-lined, indicating top and bottom of the frontal zone of inversion or relative stability which is typical of soundings through fronts.

The stratosphere is low and warm over the surface cold-air dome ("high") but high and cold over the surface low pressure. Note that the warm front connects with the tropopause in this case, but the cold air is shallow. The surface weather chart and streamlines of the tropical air flow in the warm sector above the friction layer and at the slopes of the fronts is shown in Fig. 2. Fig. 3 shows the tropopause and frontal at the same synoptic hour.

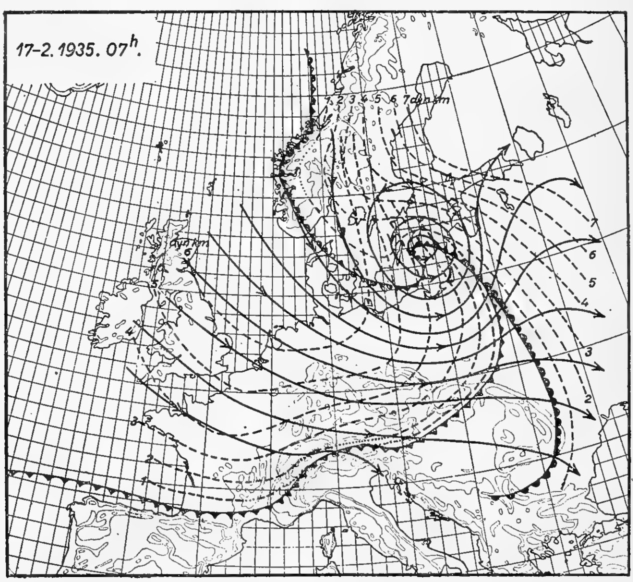


FIG. 2. Fronts and warm-air streamlines, Feb. 17, 1935, 07h. The tropical air has been far to the north over the Atlantic and is returning into the cyclone as a NW wind descending over the cold front.

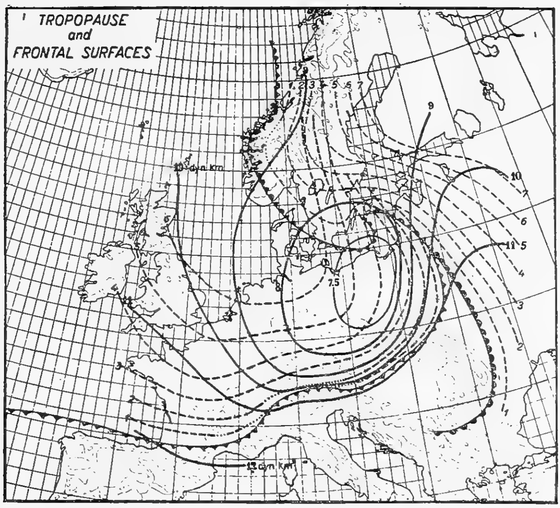


FIG. 3. The surface fronts, topography of the frontal surfaces (dashed lines) and of the main tropopause (solid lines), Feb. 17, 1935, 07h. Elevation contours for each dyn. km. The tropopause is lowest (7.5 km) just behind and south of the center of the surface low - a typical occurrence. The contours are not all open to the north so this low tropopause could not be explained as due solely to advection from the north.

The variation in the altitude of each 100 mb surface aloft as this cyclone passed over Uccle, Belgium, on 16-17 Feb., is shown in Fig. 4 (from Van Mieghem, *Ciel et Terre*, 1940, no. 1).

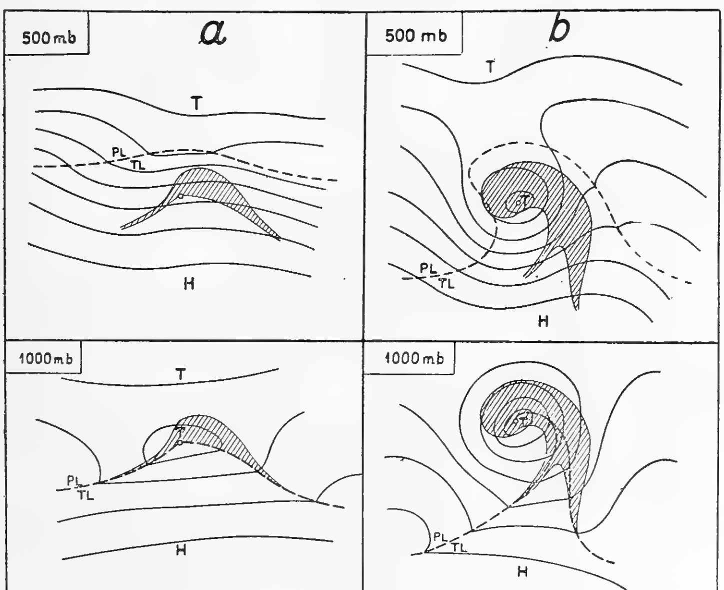
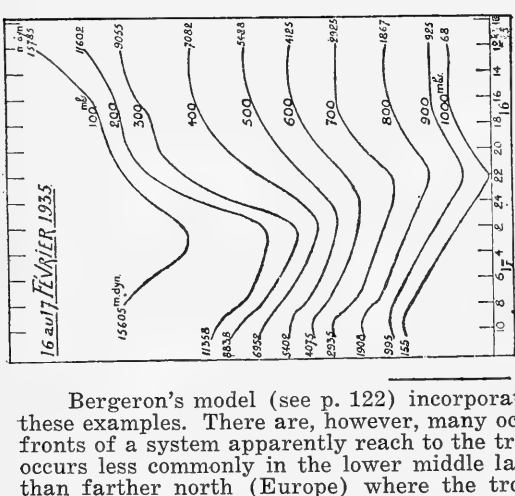


FIG. 5. Schematic streamlines in (a) a young wave-cyclone and (b) an occluded vortex cyclone, on the 500 and 1000 mb surfaces (it is customary in German weather services to draw charts of the topography of these surfaces). The shaded area is the region of frontal clouds and rain. PL = polar air; TL = tropical air; dashed lines are the fronts.



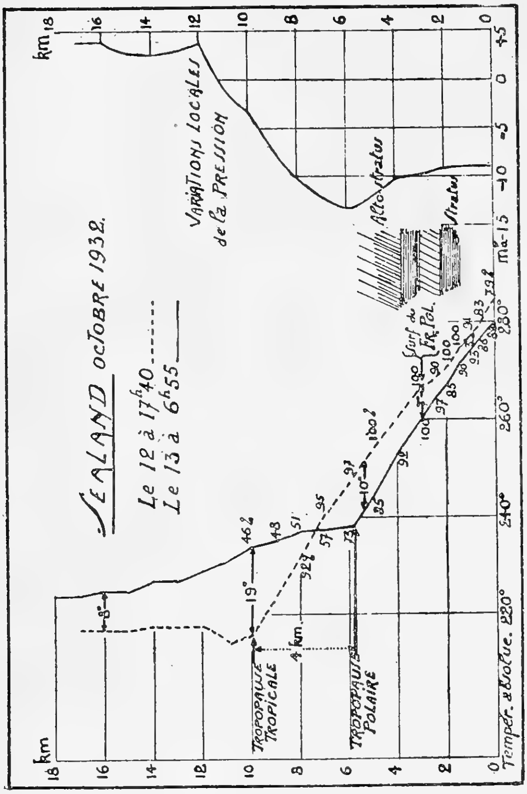
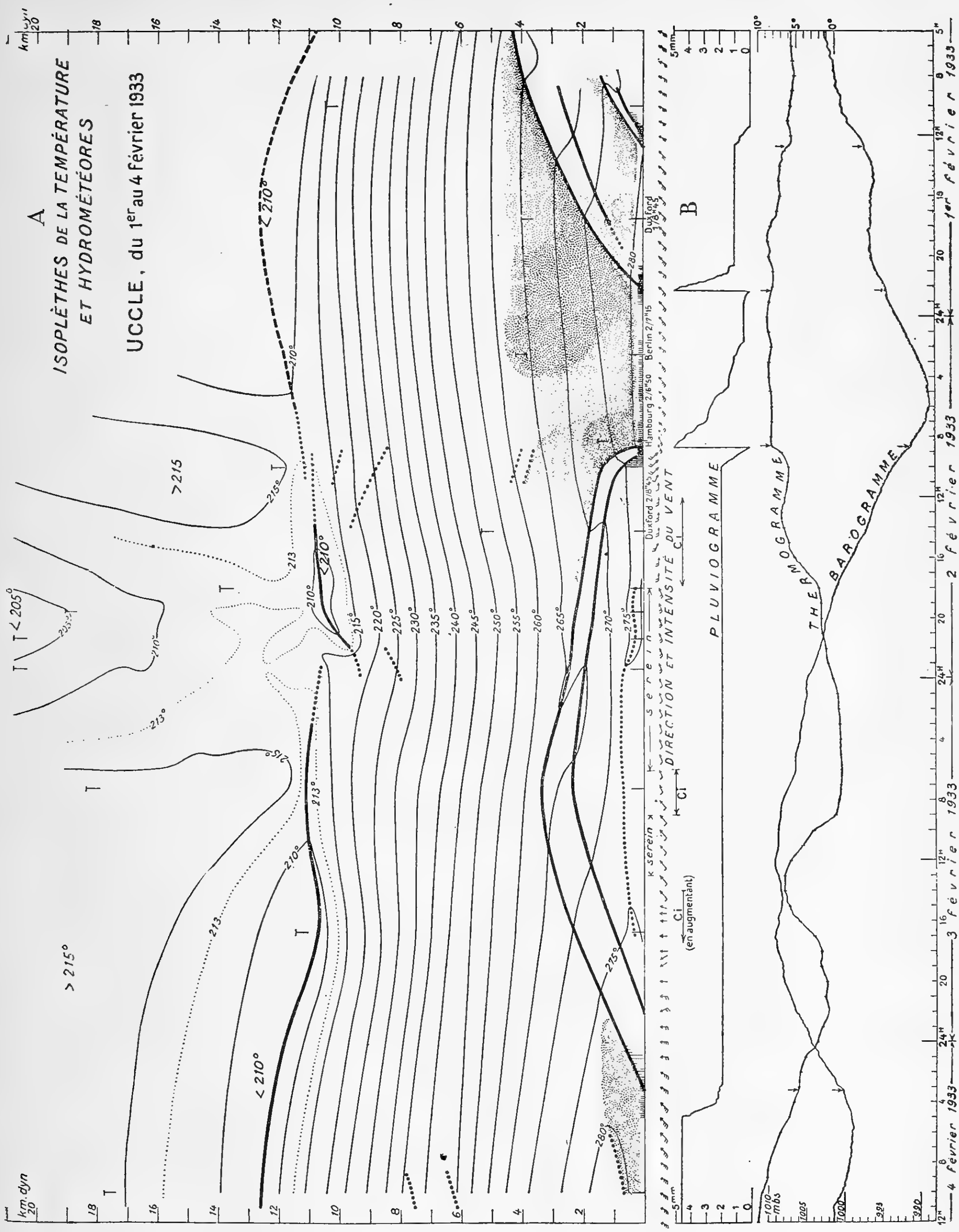


FIG. 4 (left). The lowest pressure arrived later at higher levels, as usual in cyclones; but note that the perturbation is very pronounced to over 15 km. It is of interest to inquire how much the temperature changes at each level contributed to the fall in pressure at the surface in such a case. Van Mieghem (Fig. 4, right) has done this for another storm, Oct. 12-13, 1932, which illustrates the principle. At Sealand, England, the soundings of 13 hours apart (12th and 13th) show a marked change, with the typical lowering of tropopause (by 4 km) and warming of stratosphere (by 19°C) when cold troposphere is replacing a warm one. The cold front (a long trough) had passed earlier the 12th and the cold air is 3 km deep already. The change in pressure from the 12 to 13th at each level, as drawn at the right, amounts to a maximum decrease of 13 mb at 6 km and a maximum increase of 3 mb at 12 to 16 km! The rise of temperature between 7 and 15 km was enough to cause the surface pressure to fall by 10 mb in spite of the 3° to 10°C fall in temperature of the layers below 7 km. This was of course a rapidly deepening and occluding vortex reaching to high levels. However, many of the frontal-wave cyclones and wedges or "calottes" of cold air, such as we frequently see in the daily cross-sections in United States and Europe, are rather shallow. The upper-air conditions in these are in important contrast to the deep vortex. Fig. 5 (from Bjerknes, Mildner, Palmén and Weickmann, *Ver. Geophys. Inst. Univ. Leipzig*, v. 12, no. 1, 1939, p. 61) indicates schematically this difference, as represented by streamlines drawn for the 500 and 1000 mb surfaces.

Bergeron's model (see p. 122) incorporates most of the features shown in these examples. There are, however, many occasions when both cold and warm fronts of a system apparently reach to the tropopause, though presumably this occurs less commonly in the lower middle latitudes (southern United States) than farther north (Europe) where the tropopause is normally lower. The rarity of fronts above 5 km or so is widely remarked, moreover.



In Fig. 6, shown above is a cross-section by Van Mieghem based on a series of consecutive soundings at Uccle, Belgium, and a few nearby stations, shows a very shallow cold air tongue and an open warm sector, with the accompanying clouds, surface and stratosphere weather changes. The typical multiple tropopauses and stratosphere changes are present but not nearly so pronounced as over a deep cyclone. The surface weather follows a very ideal course according to the original Bjerknes scheme of 1919. [Dotted lines are inversions of subsidence (in the troposphere) and diffuse tropopauses (at high levels); temp. in degrees Abs.; clouds stippled.] (From: *Mem. Inst. Roy. Met. Belg.*, v. 12, 1939.)

A BIBLIOGRAPHY FOR SYNOPTIC METEOROLOGISTS

By Robert G. Stone

This is by no means exhaustive but it includes most of the more important and recent papers. We give some leads to literature of the many phases of meteorology which impinge on the theory or practice of analysis and forecasting in all parts of the world. "Synoptic papers" become out of date so rapidly that only the recent literature is useful to any great extent; the earlier ones listed are either classics still of great intrinsic worth or else notable historical milestones. For English speaking countries, however, and particularly the United States, many papers of local interest are added. American synoptic meteorology emphasizes the use of upper air soundings, i.e., direct aerology, for which it is supplied with the greatest radiosonde network in the world; whereas over most of the remainder of the globe only an "indirect aerology" can be practiced. We have included representative literature in that art.

It will be understood that the authors of the various articles in this booklet do not agree with all the views one can find in these references. Furthermore, most synoptic meteorologists modify their views and practices considerably from to time in the light of new data and experience, a fact which the student should keep in mind when reading all but the latest papers.

The unorthodox classification of the literature used here is based solely on cognate fields of special interest to practical and research workers in synoptic meteorology in English speaking countries; there is no cross referencing and each title appears only once, that being under the subject for which it now appears to have the most intrinsic value, regardless of the claims of the title or intention of its author. Many theoretical papers bearing on analysis are included; popular work is omitted. Selection has been exercised to exclude most ephemeral matters in rapidly changing subjects, but some old or superficial analyses contain valuable illustrative data for the student.

Before using the bibliography study the annotated headings below.

A. GENERAL WORKS: TEXTBOOKS; HANDBOOKS; TREATISES

1. **General Meteorology (Physical and Synoptic)**
2. **Synoptic Meteorology**
3. **Dynamic Meteorology; Hydrodynamics**
4. **Special Fields of Meteorological Physics**
5. **Revolving Storms of Special Types**

Some of the more important books and papers on small vortices, waterspouts, tornadoes, whirlwind, dust devils, tropical hurricanes, typhoons, etc., from the dynamic and synoptic point of view; some papers on "dust storms" included though they are not usually vortical (see also under Section G.).

B. METEOROLOGICAL PHYSICS (RADIATION; OZONE; ICING)

Only a few of the summary and more valuable references, in which further bibliography will be found; some purely synoptic papers on aircraft icing, glaze, etc., appear under Sect. G.

C. STATICS, DYNAMICS, AND FLUID MECHANICS APPLIED TO THE ATMOSPHERE; THEORIES OF CYCLONES AND ANTICYCLONES

Theoretical papers which have either been of great influence in synoptic meteorology, or contain useful analysis of meteorological data, or which give the background for certain practical methods now current. The literature of this sort is rather thoroughly covered between the standard works by Shaw, Ertel, Bjerknes, Koschmieder, Exner, Brunt and Hann-Süring, but a selection is given here of the more important and later papers on pressure changes, perturbations, the general circulation, turbulence and energy exchange, flow patterns, particularly from the leading schools of thought now influential. More on turbulence and radiation will be found under Sections B and H. Some model experiments are cited (see also D 2, H, and J).

D. THE STATISTICAL AND SCHEMATIC BACKGROUND FOR SYNOPTIC PRACTICE:

1. Mean (General) Circulation

a. Surface Data

b. Aerological Data

In this group are chiefly statistical materials to which some interpretation is given; mere tabular results are not listed except for some remote regions where they are scarce and of special value to synoptic workers. The work by Wagner (1930) and Shaw's "Manual", vol. 2, give adequate bibliographies and summaries of the published aerological data for the world. The distinction between surface and aerological data is often arbitrary, but under (b) are found only discussions of direct upper-air observations, while under (2) appear both climatological (surface) data and indirect deductions on the upper air. Discussions of the statistics according to moving pressure systems are under D 2, but such often contain material for the general circulation, hence is is well to refer to that Section also; likewise see D 3. Also some papers under C contain incidental material for this category Ceilings, mean lapse rates, wind roses and resultants, upper mean-pressure maps, mean humidity aloft, and mean circulation patterns aloft.

2. Average Structure and Movement of Cyclones and Anticyclones; Models

The distinction between this and Sections C, G and D 1 is sometimes arbitrary; however, papers devoted primarily to statistical average conditions at the surface or aloft in cyclones and anticyclones, to the average or most frequent tracks of same (whether direct or indirect observations) are included here, while incidental information of this kind appears in many papers under the other sections mentioned. A few discussions of one or more situations selected or advocated as typical of pressure systems in general or for a given region; all schematic or model structures (some papers on typical isobaric maps appear under E), and correlations of pressure, temperature and winds aloft, find their place here. Also: structure of fronts and inversions, vorticity and solenoid distribution, symmetry points, microbaric waves, isallobaric fields, gradient winds, etc.

3. Air Mass Properties and Correlations

Statistics of air-mass properties by air-mass types, or by wind directions (see also D 1 and D 2), their frequencies, geographical ranges, sequences, effects on climate, and correlations with other geophysical phenomena and the weather type; air mass and frontal climatology.

E. SPECIAL AIDS TO ANALYSIS AND FORECASTING

1. Charts, Rules, Techniques, Thermodynamics, Formulae, Nomograms, Tables.

2. Kinematic Methods

This is a potpourri to call the attention of the student and practicing synoptic meteorologists to various aids and methods of analysis of proven or probable practical value, including diagrams, nomograms, rules for forecasting, etc. However, much ephemeral material of this sort is omitted, such as codes, obsolete principles, futile "tricks", etc. Some of the better examples of empirical statistical forecasting, indirect aerology over the oceans, and aviation problems are included; For the technique of aerological measurements and their reduction and evaluation, see the instruction handbooks of the national weather services or Linke's "Meteorologisches Taschenbuch".

F. HISTORICAL FORERUNNERS OF FRONTAL ANALYSIS

G. DETAILED ANALYSIS OF SYNOPTIC SITUATIONS

1. North and Central America

2. North Atlantic and Caribbean

3. North Pacific

4. Southern Hemisphere

5. Europe

6. Asia and Asia Minor

7. Africa (North of the Equator)

The regional and chronological divisions are for convenience primarily:—1928 marks the appearance of Bergeron's "Dreidimensional Wetteranalyse", 1937 the transition to isentropic analysis (U. S.), and 1935 to the use of divergence principles (Germany), as well as to the introduction of radiosondes, with a resulting general modification of purely Bergeronian synoptics (indirect aerology) towards direct methods. For tropical regions the papers on the general circulation (Sect. D 1) and on revolving storms (Sect. A 5) should be considered as part of the present category as well. (For mere descriptions of weather phenomena see works on general meteorology). For U. S., Germany, and Russia, where aerological data are regularly available, mostly only the studies based on more or less upper-air analysis are included.

(Outline continued, next page)

H.—METEOROLOGY OF THE FRICTION LAYER

1. **Turbulence**
2. **Surface Temperature Influences**
3. **Mountain Meteorology**
4. **Micro-Aerological Analysis and Microclimatology**

Except for some purely theoretical items in Sect. C and A3, all the papers selected dealing with the thermal and frictional influences of the earth's surface are grouped here for convenience of the many meteorologists who are specially interested in them. Otherwise these would belong to sections B, C, D, G, or I. The distinctions between "turbulence", "mountain meteorology", and "micro-aerology" cannot be made generally except from some particular point of view, which here is essentially a practical rather than a physical one. "Turbulence" covers wind structure, eddy viscosity, Austausch, evaporation, etc., over level surface or small obstructions, both in theory and in measurements to check theory or for aviation, etc. Many more references will be found in Lettau's "Atmospharische Turbulenz" and in works on Aerodynamics and Hydrodynamics (Sect. A3). "Mountain Meteorology" covers large-scale turbulence and special winds due to hills and mountains, as well as the departures of mountain observations from free-air conditions and the adaptation of mountain observations in synoptic or dynamic meteorology. "Micro-aerology" includes the actual observations of air structure (synoptic and average) in layers close to the ground to show surface influences, radiation, etc. Turbulence measurements are chiefly listed under that heading, however. "Micro-climatology" provides data on representativeness of surface weather reports and the mean circulation in the lowest layers; it is a valuable link in the application of synoptic meteoroogy to many practical ploblems.

I. PRECIPITATION AND CONDENSATION; CONVECTION; CLOUDS; STRATUS AND FOG

No attempt has been made to separate these topics as they intergrade and the number of citations is not inconveniently great. Only recent and summary papers on condensation and precipitation theory are cited (see also Sect. B., and A 4). "Convection" is here restricted to vertical convection from surface heating, and convergence, lifting, etc., resulting in instability and clouds, with applications to gliding, aviation, weather analysis and forecasting. The theory of lapse rates and of the kinetic energy of thermal stratifications is partly covered in Sections A, C, and E. Of the large literature on clouds some of the best aerological studies and synoptic applications have been selected; much further valuable material of this sort will be found incidentally in Sections B, D, E, G, and notably Süring's book gives further bibliography, especially on other aspects of clouds.

J. INSTRUMENTAL PROBLEMS AND DEVICES

Improvements in aerological instruments come so rapidly that only late work is of much practical value except to specialists in instruments. The instruction handbooks issued by the weather services of each country should be consulted for the national and official practices, which vary greatly. See also Sect. E 1, B, and C.

Abbreviations

The following condensed abbreviations are used. Other abbreviations are more complete and will probably be recognized at sight.

Ann. d. Hydr.	Annalen der Hydrographie und maritimen Meteorologie, Berlin.
BAMS	Bulletin of the American Meteorological Society, Milton, Mass.
Beitr. Phys. fr. At.	Beiträge zur Physik der freien Atmosphäre, Leipzig.
Geog. Ann.	Geografiska Annaler, Stockholm.
Geofys. Publ.	Geofysiske Publikasjoner, Norske Videnskaps Akademie, Oslo.
Gerl. Beitr. Geophys.	Gerlands Beiträge zur Geophysik, Leipzig.
Met. Mag.	Meteorological Magazine, London.
MWR, or M. W. R.	Monthly Weather Review, Washington.
MZ, or M. Z.	Meteorologische Zeitschrift, Braunschweig.
QJRMS, or Q. J. R. M. S.	Quarterly Journal of the Royal Meteorological Society, London.
UGGI	Union Géodésique et Géophysique Internationale.

A. GENERAL WORKS; TEXTBOOKS, HANDBOOKS, TREATISES

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BERGERON, T., *The Physics of Fronts* (English abstract of Pt. II of "über die dreidimensional Verkn. Wetteranalyse," 1928), planographed, the author, 1936 full paper to appear later in Geophys, Publ.] Reprinted in BAMS, v. 18, pp. 265-75, 1937, and in *Das Wetter,* Dec. 1936.

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Glossary of Elementary Terms Used in Articles I to IX

absolute humidity (II)[1]—the mass of water vapor present in a unit volume of air, or the density of the water vapor.

absolutely stable (I)—a vertical distribution of temperature such that, whether the air be dry or saturated, particles will tend to remain at their original level. In this case the lapse rate must be less than the saturation adiabat at the prevailing temperatures.

absolutely unstable (I)—a lapse rate greater than the dry adiabatic; both dry and saturated air are unstable. Also called a *super-adiabatic lapse rate.*

adiabat (I)—a curve along which a thermodynamic change takes place without the addition or subtraction of heat. In the case of the atmosphere a *dry adiabat* is generally considered a temperature-height or temperature-pressure curve along which a rising or sinking air particle will fall providing no saturation occurs and providing, of course, that no heat is given to or taken from the particle in its path. Similarly a *wet adiabat* (*saturation adiabat, condensation adiabat,* or *pseudo-adiabat*) is a temperature-height or temperature-pressure curve along which the saturated rising particle will fall.

adiabatic chart (III)—a thermodynamic diagram in which temperature is plotted against pressure (either on a logarithmic scale, or pressure to the 0.288 power) and in which dry adiabats are constructed. The chief use of this chart is for evaluation of aerological soundings.

adiabatic process (I)—a thermodynamic process in which no heat is transferred from the working substance to the exterior or vice versa; a thermally insulated process.

adiabatic rate of cooling with ascent for dry air (I)—very nearly constant in the troposphere at 1 degree Centigrade per 100 meters (see *adiabat*).

adiabatic rate of cooling with ascent for saturated air (I)—a rate which varies chiefly with the temperature and hence has no fixed value.

air mass (II)—an extensive body of air which approximates horizontal homogeneity.

characteristic curve (IV)—the curve joining the significant points of an aerological sounding when plotted on the Rossby diagram.

cold front (V and VII)—the discontinuity in front of a wedge of cold air which is displacing warmer air in its path.

conditional equilibrium (I)—a vertical distribution of temperature such that the layer is stable for dry air but unstable for saturated air. In this case the lapse rate lies between the dry and the saturated adiabat. Also called *conditional instability,* or *conditional stability,* and *moist labile equilibrium.* (See *latent instability.*)

conservatism (II)—the degree of constancy of a meteorological element when the given air mass is subject to modifying factors.

convective instability (IV)—a vertical distribution of temperature and moisture such that lifting of the entire layer will eventually lead to instability with respect to dry air. In convective instability the equivalent-potential temperature decreases with elevation. Also called *"potential instability". "Convective equilibrium,"* however, merely means an unstable (adiabatic) lapse rate.

depegram (VIII)—a curve representing the behavior of the dew-point with pressure changes for a given sounding, drawn on the tephigram.

discontinuity (I)—a zone of comparatively rapid transition of the meteorological elements. These discontinuities are not mathematically abrupt, but are rapid transitions

[1]Roman numerals following each term refer to the article in which the topic is discussed in some detail.

compared with the ordinary transitions in one and the same air mass. (Practically synonymous with *front.*)

dry air (I)—air which is not saturated.

emagram (VIII, IX)—any adiabatic diagram with coordinates T, log p. Areas are proportional to energy as on the tephigram.

energy diagram (III, VIII, IX)—any thermodynamic diagram on which area is proportional to energy; such as the *tephigram* or *emagram* (which see).

equivalent-potential temperature (II)—the temperature a given air particle would have if it were brought adiabatically to the top of the atmosphere (i.e., to zero pressure) so that along its route all the moisture were condensed (and precipitated), the latent heat of condensation being given to the air, and then the remaining dry sample of air compressed adiabatically to a pressure of 1000 millibars.

equivalent-potential temperature diagram—see *Rossby diagram.*

equivalent temperature (II)—the temperature a particle of air would have if it were made to rise adiabatically to the top of the atmosphere (i. e., to zero pressure) in such a manner that all the heat of condensation of the water vapor were added to the air and the sample of dry air were then brought back adiabatically to its original pressure.

estegram (VIII)—the characteristic curve of wet bulb temperatures of a sounding plotted on a tephigram or pseudo-adiabatic chart.

front (V, VI and VII)—the discontinuity between two juxtaposed currents of air possessing different densities. Most frequently fronts represent the boundary between different air masses. The so-called "*wind-shift line*" is usually a well-marked front.

frontogenesis (VII)—the creation of fronts generally brought about through the horizontal convergence of air currents possessing widely different properties.

frontolysis (VII)—the destruction of fronts generally brought about by horizontal divergence at the discontinuity zone.

instability (I)—the opposite of *stability;* a lapse rate in which particles will be readily displaced vertically upon small impulse. (Also called *lability.*) See: *conditional, latent, pseudo-, mechanical* and *absolute instability.*

instability showers (V)—showers caused by steepening of the lapse-rate in any way, such as the rapid warming of the lower layers of a cold current as it moves over a relatively warm surface. In most cases there is an appreciable addition of moisture to the lower layers, as for example, when a polar continental current moves over a body of warm water.

lapse rate (I)—the existing rate of change of an element, commonly temperature, with height in a given layer of the atmosphere.

latent instability (VIII)—on the tephigram, when the area of positive energy is greater than the negative energy area (this is more properly called *real latent instability*). In general *latent instability* refers to the energy that can be released after the convection reaches the condensation level. It is the case of *conditional instability* where the air is moist enough for convection to form clouds. (See also *pseudo-instability.*)

loop (*or bent*) *back occlusion* (VII)—an occluded front which has bent back in the rear of the cyclone so that it appears in the meteorological field as another front behind the cold front. In most cases these occlusions are of the cold front type. That is, the air behind is colder than that preceding them.

mechanical instability (I)—a lapse rate such that the air density decreases with elevation; for this condition the lapse rate must be greater than 3.42 degrees C. per 100 m; also called *"auto-convection"*, since no initial impulse is needed to set it off; *"self-starting"*.

mixing ratio (III)—the mass of water vapor per unit mass of perfectly dry (absence of water vapor) air.

$w = 622\, e/(p—e)$ grams per kilogram (see *symbols*)

modification of air mass properties (II)—the change in the values of the meteorological elements within an air mass due to such influences as radiation, turbulence, subsidence, convergence, and so on. These modifying influences tend to destroy the original horizontal homogeneity of the air mass.

negative area (VIII)—the area on a tephigram enclosed between the path of the rising particle and the surrounding air when the rising particle is at every stage in its ascent colder than the environment.

neutral equilibrium (I)—a vertical distribution of temperature such that a particle of air displaced from its level neither assists nor resists the displacement; that is, at every level the density of the displaced particle is equal to that of the surrounding air. In the case of dry air the corresponding lapse rate is that of the dry adiabat; in the case of saturated air, the saturation adiabat.

occluded front (VII) or *occlusion*—the front formed when and where the cold front overtakes the warm front of a cyclone. This front marks the position of an upper trough of warm air, originally from the warm sector, which has been forced aloft by the action of the converging cold and warm fronts. Occlusions may be of the warm front type in which the air in advance of the front is colder than that behind, or of the cold front type, in which the air in advance is the warmer.

Occlusion is also the term used to denote the process whereby the warm air of the cyclone is forced from the surface to higher levels.

partial potential temperature (III)—the temperature a given air particle would have if it were reduced adiabatically from the pressure exerted solely by the dry air to a pressure of 1000 mb.

$\Theta_d = T\,[1000/(p—e)]^{0.288}$

(see *symbols*)

penetrative convection (IV)—small convective up-currents locally penetrating an overlying more stable layer without generally or greatly altering the existing atmospheric stratification.

Poisson's equation (I)—the relation between temperature and pressure in dry air which is undergoing adiabatic transformation.

$T_1/T_2 = (p_1/p_2)^{0.288}$

polar front (VII)—the frontal zone between air masses of polar and those of tropical origin.

positive area (VIII)—the area on a tephigram enclosed between the path of the rising particle and the surrounding air when the rising particle is at every stage in its ascent warmer than the environment.

potential temperature (II)—the temperature a given particle of air would have if it were reduced adiabatically to a pressure of 1000 mb.

$\Theta = T(1000/p)^{0.288}$

pseudo-adiabatic (IV)—the process wherein a saturated air particle undergoes adiabatic transformations, the liquid water being assumed to fall out as it is condensed.

pseudo-adiabatic chart (III, VIII)—an *adiabatic chart* on which wet adiabats are also drawn; lines of saturation specific humidity are usually added too. Used for analyzing stability conditions in a sounding (see *emagram*).

pseudo-(latent) instability (VIII)—on the tephigram, when the area of positive energy is less than the area of negative energy.

relative humidity (II)—the ratio of the actual vapor pressure and the maximum vapor pressure possible at the same temperature.

$f = e/e_m$ (see *symbols*)

representative observations (II)—those which give the true or typical conditions of the air mass; hence they must be relatively uninfluenced by local conditions and taken from outside the transition zones and fronts.

Rossby diagram (III)—a thermodynamic diagram making use of the highly conservative air mass properties: partial potential temperature, equivalent-potential temperature and mixing ratio.

secondary fronts (VI)—fronts which develop at some distance from the principal fronts of the cyclone. These fronts are often the result of dynamic effects behind the cold front, or are merely loop back occlusions.

slope of a front (V)—the tangent of the angle formed by the discontinuity surface and a horizontal plane.

source region (II)—an extensive area of the earth's surface characterized by sufficiently uniform surface conditions and which is so placed in respect to general circulation that masses of air may remain over them sufficiently long to take on fairly definite properties.

specific humidity (II)—the mass of water vapor in a unit mass of moist air. $q=622\ e/p$ grams per kilogram.

squall head (VI)—the piled up cold air at the cold front, sometimes taking the form of an overhanging tongue. Also *"line squall"*.

stability (I)—a vertical distribution of temperature such that particles will resist displacement from their level. In the case of dry air the lapse rate for stability will be less than the dry adiabat; in that of saturated air, less than the saturation adiabat.

stratification (V)—a layering of the atmosphere, so that each layer is characterized by a particular temperature distribution and moisture content. Instability tends to wipe out stratification as it brings about mixing.

subsidence (IV)—an extensive sinking process most frequently observed in polar anticyclones. The subsiding air is dynamically warmed and made more stable.

superadiabatic (I, VIII, IX)—a lapse-rate greater than the dry adiabatic; absolute instability. Mechanical instability may be implied also.

surface of discontinuity (V)—the sloping boundary zone between air masses of different properties (see *discontinuity*).

symbols—used throughout the series and in the formulas given herein:

e =vapor pressure
e_m =saturation vapor pressure
f =relative humidity
p =total pressure
q =specific humidity
t, t'=dry, and wet, bulb temperature, resp., in °F or in °C
T=absolute temperature
T'=wet-bulb temperature (°A)
w =mixing ratio
θ =potential temperature
T_E=equivalent temperature
θ_E =equivalent-potential temperature
θ_d =partial potential temperature
θ'=wet-bulb potential temperature.
ϕ =entropy

tephigram (VIII)—a thermodynamic diagram for estimating the quantity of available convective energy in the overlying air column; also applied to the graph of an individual sounding plotted with coördinates temperature and entropy.

transition zone (V)—the zone at a discontinuity wherein the properties are characteristic neither of one air mass nor the other, but lie somewhere between the two. It is now customary to assume that all the air in the transition zone belongs to the colder air mass, the air in warm sectors being considered more nearly homogeneous.

unstable (I)—a vertical distribution of temperature such that particles of air, because of their lesser or greater density than the surrounding air, will rise or sink of their own accord once given an initial impetus up or down. For dry air the unstable lapse rate is greater than the dry adiabat; in the case of saturated air, greater than the saturation adiabat.

upper front (VII)—a front whose principal development and evidence is in the upper air, usually the active *upper cold front* of a warm front type occlusion.

vapor pressure (II)—the partial pressure of the air exerted solely by the water vapor molecules.

warm front (V)—the discontinuity at the front of a warmer air mass which is displacing a retreating colder air mass.

warm sector (VII)—the air enclosed between the cold and warm fronts of a cyclone.

wave disturbance (VII)—a deformation produced along a front. These waves travel along the discontinuity surface often producing new cyclones.

wet-bulb temperature (III, VIII)—the lowest temperature to which a wetted ventilated thermometer (as of a psychrometer, e.g.) can be cooled, i.e., by evaporation. while not strictly a temperature of the air, it is a function of dry bulb, relative humidity and barometric pressure. On the adiabatic chart or tephigram: find the intersection of the dry adiabat and specific humidity line and then follow down the wet adiabat to the original pressure level of the point in question. (See: *estegram; latent instability; pseudo-instability.*) The wet adiabats are isotherms of equal *wet-bulb potential temperature* (also equivalent-potential temperature).

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