INTRODUCTION
TO
METEOROLOGY

BY

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INTRODUCTION TO METEOROLOGY

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Preface

The aim of this book is to present in an elementary manner the basic principles of modern meteorology. Intended for students without previous acquaintance with the subject, it is written in response to a demand for a nontechnical text to serve the many short and elementary courses in meteorology already in progress in the United States. It is written more for the purpose of creating interest and background than for furnishing a technical and detailed discussion of the various branches of meteorology. For this reason the use of calculus in the presentation of meteorological theories has been avoided.

Actually, this book is an expansion of a chapter on meteorology written for the British Empire edition and the American edition of Weems's "Air Navigation" and an abbreviation of the author's recent book "Weather Analysis and Forecasting." However, this book is not addressed specifically to pilots or weather forecasters, although its leaning is decidedly toward synoptic and aeronautical meteorology.

On account of its elementary character, the text does not contain references to meteorological journals and papers. Instead, a list of advanced textbooks is included to assist the reader in finding more advanced literature on meteorology.

The author's warm thanks go to Miss Margaret Whitcomb for her valuable assistance in revising the text, preparing the diagrams and tables, proofreading, and indexing.

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INTRODUCTION

Meteorology is the science of the atmosphere. With the increasing tendency toward specialization characteristic of our time, the subject matter under the general heading of meteorology may be referred to various subdivisions, or branches, depending partly on the theoretical approach and partly on the application of meteorology to human activities.

From a theoretical point of view, meteorology may be subdivided into the following categories:

1. *Dynamic meteorology,* which concerns itself with the forces that create and maintain motion and the heat transformations associated therewith. Within the field of dynamic meteorology, distinction is often made between *hydrodynamics* which deals with forces and motion and *thermodynamics* which deals with heat. The word *aerodynamics* is usually reserved for the study of the interaction between air currents and objects, such as airfoils.

2. *Physical meteorology,* which deals with processes of a purely physical nature, such as radiation, heat, evaporation, condensation, precipitation, ice accretion, and optical, acoustical and electrical phenomena.

3. *Climatology,* or statistical meteorology, which determines the statistical relations, mean values, normals, frequencies, variations, distribution, etc., of the meteorological elements.

From the point of view of practical application, meteorology is commonly subdivided into a number of classes, of which the following are the most important:

4. *Synoptic meteorology* has as its aim a coordinated study of the processes in the atmosphere on the basis of simultaneous observations over large areas. Thus, synoptic meteorology applies

dynamic as well as physical meteorology and, to a lesser extent, climatology in order to obtain a synopsis of the state of the atmosphere; its main purpose is the analysis and forecasting of the weather phenomena.

5. *Aeronautical meteorology* deals with the application of meteorology to the problems of aviation. As far as actual weather conditions are concerned, it is related to synoptic meteorology; and as far as the normal state of the atmosphere is concerned, it is related to climatology.

6. *Maritime meteorology* is related to marine navigation in the same manner as aeronautical meteorology is related to air navigation.

7. *Agricultural meteorology* deals with the applications of meteorology to agriculture, soil conservation, etc.

8. *Hydrometeorology* is concerned with meteorological problems relating to water supply, flood control, irrigation, etc.

9. *Medical meteorology* has to do with the influence of weather and climate on the human body.

10. *Aerology* is the branch of meteorology that is concerned with the conditions of the free atmosphere on the basis of direct observations. In the United States the word aerology is frequently used synonymously with meteorology, meaning the science of the atmosphere.

The scope of the present book is mainly synoptic, inasmuch as it is chiefly concerned with the weather processes; in addition, a brief chapter on climates and an outline of the history of meteorology are included.

CHAPTER I

THE ATMOSPHERE

The word atmosphere derives from the Greek words "atmos" which means vapor and "sphaira" which means sphere. It is now used to denote the gaseous sphere surrounding the earth.

Composition.—The air, or the material of which the atmosphere is composed, is a mechanical mixture of a number of different gases. A sample of dry and pure air contains about 78 per cent (by volume) nitrogen, 21 per cent oxygen, and almost 1 per cent argon. In addition, it contains about 0.03 per cent carbon dioxide.

Nitrogen, oxygen, argon, and carbon dioxide constitute about 99.99 per cent of dry and pure air. The remaining 0.01 per cent represents traces of several other gases, such as neon, krypton, helium, ozone, xenon, and hydrogen. These are present in such minute amounts that they are of no practical importance for the study of the weather phenomena.

The amount of carbon dioxide is not quite constant. The vegetable world continuously consumes carbon dioxide which, again, is produced by the animal world, through burning of fuels, volcanic action, and various processes of decay in the soil. Although these processes are not always balanced, the oceans, by dissolving the excess of carbon dioxide, so effectively regulate it that the amount of carbon dioxide in the atmosphere remains almost constant.

Ozone, which is present in the lower atmosphere in minute amounts, has a maximum in the upper atmosphere between 10 and 25 km. (30,000 and 80,000 ft.) where its amount varies considerably.

Apart from the variations in carbon dioxide and ozone, the composition of the atmosphere is remarkably constant all over the earth's surface. It is also constant with elevation as far as instruments have reached (35 km.).

The air also contains a variable amount of water vapor. In many respects the water vapor is the most important constituent of the atmosphere. The maximum amount of water vapor that the air can absorb depends entirely on the temperature of the air; the higher the temperature of the air, the more water vapor can it hold. The air is saturated with moisture when this maximum amount is reached. When air is cooled below its saturation temperature, condensation takes place, the water vapor being condensed to water droplets or, at low temperatures, to ice crystals. Small water droplets and ice crystals are kept afloat in the air by the ascending air currents. Under special conditions, which we shall describe later, these minute drops or ice crystals coalesce and form large drops or snowflakes which are precipitated from the clouds when they become too large to be kept afloat by the ascending currents.

Impurities.—Apart from the above-mentioned constituents, the air contains a variable amount of impurities, such as dust, soot, and salts.

The main source of dust is the arid regions, such as deserts and steppes. The coarser material, whirled up by the winds, is never carried far from its source, but minute dust particles are readily distributed throughout the lower atmosphere and carried far from the source. Air masses that have swept over subtropical continents normally contain considerable amounts of dust, but polar air masses are relatively pure.

The industrial regions, forest fires, and volcanoes constitute the main source of soot. When fuels burn at high temperatures, hydrogen and oxygen combine to produce water vapor, and carbon and oxygen combine to produce carbon dioxide, both of which belong to the normal constituents of the atmosphere. However, if fuels burn at low temperatures, the latter process is hindered, and carbon is carried up with the rising air and coagulates into soot.

Observations show that the air normally contains a considerable amount of salts. Through the action of the winds, spray is whirled up from the oceans, and when it evaporates the salt remains in the air in the form of minute particles.

The particles which constitute the impurities of the air are so small that they cannot be seen individually by the naked eye, but their effect on visibility and on coloring of distant objects is easily observed. Through haze, distant objects (*e.g.*, mountains) are seen as if through a thin veil of pale blue if the object is dark or a yellowish veil if the object is white (*e.g.*, snow-covered mountains, clouds at the horizon). At a certain distance depending on the density of the haze, all details disappear, and the objects stand out like a silhouette against the sky. The denser the haze, the shorter the distance at which the details disappear.

The presence of dust in the atmosphere is important not only because of its influence on visibility; if the air were perfectly pure, there would be no appreciable condensation of water vapor. When the air is cooled to its saturation temperature, condensation takes place on certain active (hygroscopic) nuclei. Salt particles from the oceans and various products of combustion are most active as condensation nuclei, and observations show that such particles are present in the atmosphere in abundant amounts.

Structure.—The air is highly elastic and compressible. Although extremely light, it has a perfectly definite weight. At ordinary pressure and temperature, the weight of a sample of air near the earth's surface is about $\frac{1}{800}$ of the weight of an equal volume of water. Thus, 1 cu. ft. of air weighs about 1.2 oz. or, in metric units, 1 cu. m. of air weighs about 1.3 kg.

In consequence of this weight, the atmosphere exerts a certain pressure upon the earth's surface, amounting to about 15 lb./sq. in. This pressure is sometimes used as a unit and is called 1 atmosphere.

A column of air from the earth's surface to the top of the atmosphere exerts a pressure on the earth's surface that is equivalent tc a column of water 34 ft. (or 10 m.) high; this is equivalent to the weight of a column of mercury 30 in. (or 76 cm.) high. For this reason, mercurial barometers are used to measure the pressure of the atmosphere.

The atmospheric pressure decreases with increasing altitude. The difference in pressure between two points, one above the other, is simply equal to the weight of the air column between the two points. By measuring the temperature and the pressure at two or more points in the same vertical line, it is possible to compute the difference in altitude between the points. Assuming a normal distribution of temperature, pressure becomes a simple function of altitude, and special barometers (*altimeters*) have been constructed to record altitude instead of pressure. Such instruments are now widely used in aviation.

Since the atmospheric pressure is equal to the weight of the air column, it follows that the pressure must decrease gradually and approach zero with increasing altitude. The same also applies to the density of the air. There is, therefore, no distinct upper limit to the atmosphere; it merges gradually into empty space. Figure 1 shows how pressure, density, and temperature normally vary with altitude within the lower 15 km.

Even though the atmosphere reaches to great altitudes, it is only the lower part of it that is of importance for the weather.

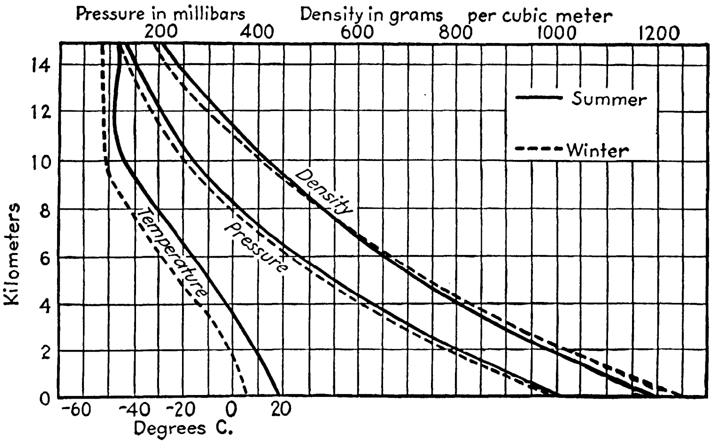


FIG. 1.—Showing the normal variation in temperature, pressure, and density with height in middle latitudes.

The highest clouds (cirri) are seldom more than 10 km. (33,000 ft.) above the earth's surface, and 50 per cent of the total weight and about 90 per cent of the total moisture content are within about 5 km. (16,000 ft.) of the earth's surface.

Stratification.—It will be seen from Fig. 1 that the air temperature normally decreases with elevation up to about 11 km. (36,000 ft.) and then remains constant. The rate of decrease in temperature along the vertical is called the *lapse rate*. The lower part of the atmosphere, which normally is characterized by a relatively steep lapse rate, is called the *troposphere*. The upper part of the atmosphere, which is characterized by almost constant temperature along the vertical, is called the *stratosphere*. The layer of transition that separates the stratosphere from the troposphere is called the *tropopause*.

The height of the tropopause above the earth's surface varies considerably with latitude and season. It also varies with the weather situation; it is normally lower over areas of low pressure (cyclones) than over areas of high pressure (anticyclones).

Figure 2 shows the normal height of the tropopause and the mean distribution of temperature in the lower atmosphere. It is of interest to note that the temperature of the stratosphere, on the whole, decreases from the poles toward the equator.

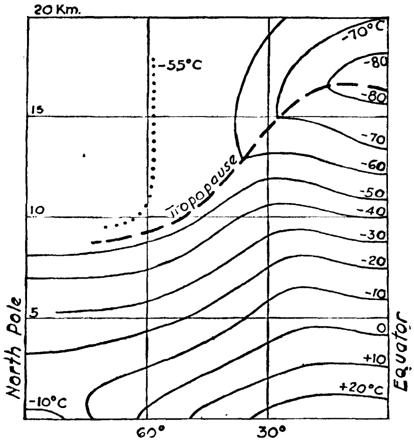


FIG. 2.—Mean annual temperature in the troposphere and the lower stratosphere. Note that the stratosphere is warmer at the pole than at the equator.

The greatest altitude to which meteorological instruments have reached is about 36 km. From the direct observations made below 36 km. and from recent studies of radiation, meteors, aurora borealis, the propagation of sound and radio waves, etc., it is possible to draw conclusions as to the structure of the upper atmosphere. Our present knowledge of the stratification of the atmosphere may be summarized briefly as follows:

In the troposphere the temperature normally decreases with altitude at a rate of approximately 0.6°C. per 100 m., or about 1°F. per 300 ft. The troposphere is relatively unstable; vertical currents occur frequently, leading to condensation, the formation of clouds and precipitation. All ordinary weather phenomena develop within the troposphere, particularly in its lower half.

As we ascend through the tropopause (see Fig. 3), the temperature is constant or increases along the vertical as far as meteor-

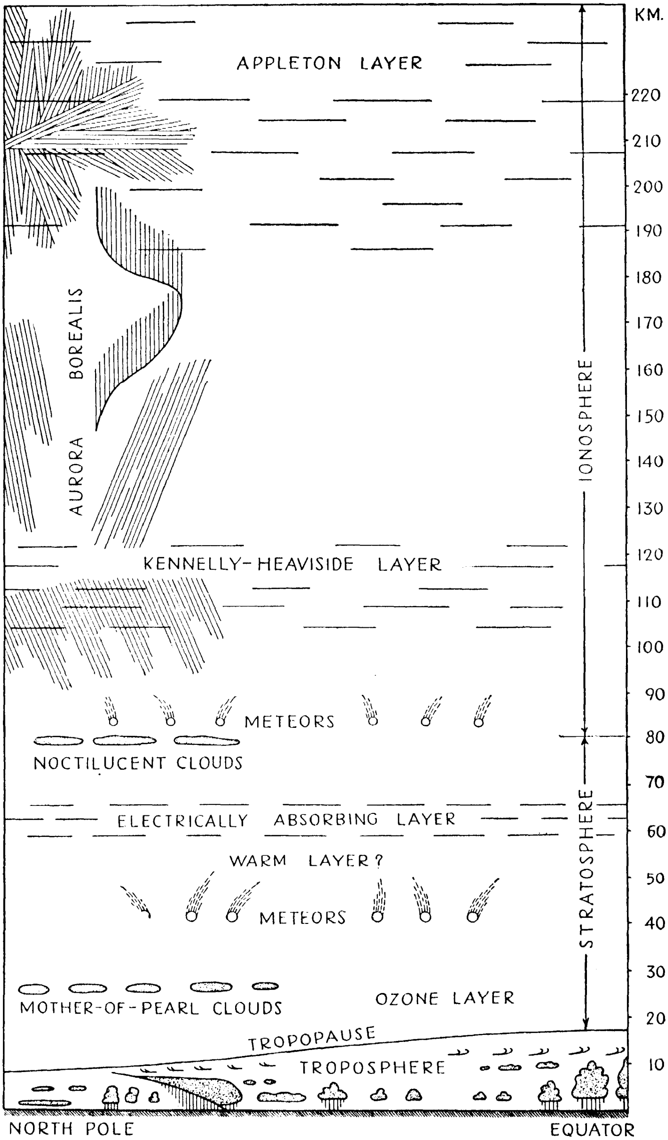


FIG. 3.—Showing the structure of the troposphere, stratosphere, and ionosphere and the phenomena typical of each.

ological instruments have reached. Above the tropopause, we meet with a layer particularly rich in ozone. Recent investigations by Dobson and others have shown that there is a noticeable correlation between the amount of ozone and the weather conditions at the ground. Although the stratosphere usually is cloudless, a special type of cloud (mother-of-pearl clouds) is occasionally observed to form in connection with the ozone layer. The space between the tropopause and the ozone layer (see Fig. 3) is always cloudless; and since the air within this layer is extremely stable, the lower stratosphere offers the nearest approach to ideal flying conditions.

Statistical investigations show that the meteors disappear most frequently either at about 80 km. (260,000 ft.) or at about 40 km. (130,000 ft.) above the earth's surface. This fact and the results of the study of the propagation of sound waves seem to indicate that there is a layer of air between 40 and 80 km. which is extremely warm, perhaps 120 to 130°F. (*i.e.*, 60 to 70°C.).

At about 60 km. (200,000 ft.), there is a layer that tends to absorb radio waves. This layer is created through the action of the sun's rays; and, as a result, the range of the radio stations, particularly that of short-wave stations, is greater by night than by day.

Above the level of about 80 km. (260,000 ft.) is the so-called *ionosphere.* In the lower portion of it, we find the noctilucent clouds, which are extremely rare. The ionosphere is characterized by several electrically conducting layers, of which the *Kennelly-Heaviside layer* (or the *E*-layer) is the most important. This layer reflects the radio waves back to the earth's surface and thus accounts for the long range of radio stations. The Kennelly-Heaviside layer, which is quite distinct, is normally found between 90 and 130 km. (300,000 to 400,000 ft.). Above this layer is the so-called *Appleton layer* (or the *F*-layer), which is more diffuse and variable in altitude; it sometimes breaks up into several diffuse layers.

The aurora borealis and kindred phenomena are most frequently observed within the lower part of the ionosphere. Recent measurements of Störmer have shown that auroras may occur even as high as 1200 km. (4 million feet) above the earth's surface. This shows that atmospheric matter is present in measurable amounts even at such great altitudes.

CHAPTER II

OBSERVATIONS AND INSTRUMENTS

The meteorological services of the various countries maintain a large number of observing stations whose function it is to make observations according to international definitions and rules and to report at frequent intervals to the central offices. Within each country the reports are collected and disseminated by means of radio, teletype, or telegraph. By international agreements, all countries are obliged to arrange for radio transmission of meteorological reports for international use. All matters of international importance are handled by the International Meteorological Organization, which maintains a permanent secretariat in Lausanne, Switzerland. Meteorological observations are also made by a large number of selected ships whose reports are exchanged among the various countries.

The meteorological observing stations may be divided into three groups, *viz.:*

1. Ordinary land and ship stations which report the conditions near the earth's surface and the state of the sky.
2. Pilot-balloon stations which measure the winds in the free atmosphere.
3. Aerological stations which send up balloons or airplanes furnished with instruments to measure pressure, temperature, and humidity in the free atmosphere.

In this chapter, we shall consider the most important meteorological elements and the instruments commonly used. The reader who desires more detailed information is referred to the publications issued by the national weather services.

The Mercurial Barometer.—The atmospheric pressure is usually observed by means of a mercurial barometer. The principle of the mercurial barometer may be described briefly as follows: A glass tube, about 1 yd. long, with one end sealed and the other end open, is filled completely with mercury. The open end is now closed (*e.g.*, by a finger), and the tube is placed in a vertical position with the open end submerged in a small vessel partly filled with mercury (Fig. 4). When the finger is removed, the mercury in the tube will sink somewhat and come to rest at a level of about 30 in. above the level of the mercury in the vessel. There is then a vacuum above the mercury in the tube and, therefore, no atmospheric pressure above the mercury within the tube.

Since the atmospheric pressure acts on the free surface of the mercury in the vessel, it is clear that the weight of the mercury column above the free surface of the mercury in the vessel must be equal to the weight of the air column above the same surface. The length of the mercury column will at any moment indicate the atmospheric pressure; it is measured by means of a scale placed along the glass tube.

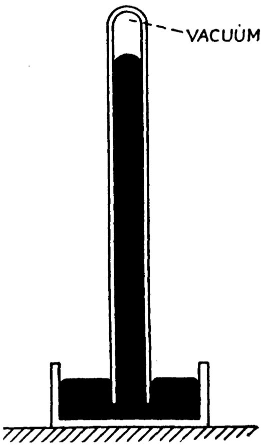


FIG. 4.—Showing the principle of the mercurial barometer. The air pressure on the mercury in the vessel is balanced by the weight of the mercury in the tube.

Pressure.—The direct reading of the barometer gives the *length* of the mercury column whose *weight* balances the weight of the air column above the barometer. The length of the column of mercury depends on the temperature of the barometer. In order to render the observations of the various stations comparable, the readings are corrected for the thermal expansion of the tube, the mercury and the scale, using 0°C. (32°F.) as standard temperature.

The weight of the air column depends also on the local gravity. A second correction is applied to obtain the pressure that would be observed at the station if the local gravity were equal to the normal gravity at 45°N.

Even the most carefully made barometers have a certain error, and the reading must be corrected for the individual error of the instrument. When the reading has been corrected for instrument error, thermal expansion, and local gravity, we obtain the correct pressure at the level of the mercury cistern of the barometer.

The atmospheric pressure decreases with elevation at a rapid rate. In order to render the observations of the various stations comparable, it is necessary to add a correction so as to obtain the pressure that would be recorded if the barometer were placed at sea level. Hence, the pressure reported from all land stations and ships indicates the pressure that would be recorded by a correct barometer if it were placed at sea level, if its temperature were 0°C. (32°F.), and if the local gravity were equal to the normal gravity at 45°N.

Pressure Units.—Up to about 1914, pressure was reported in units of length, either in inches of mercury or millimeters of mercury. In later years a new unit, called the millibar (mb.) has come into general use. Normal pressure at sea level is roughly 30 in., or 760 mm., to which corresponds 1013 mb.

The conversion from units of length to units of pressure proceeds as follows: Suppose that the mercury column is 76 cm. high and of unit cross section. Since the density of mercury is about 13.595, the mass of the column of mercury would be $76 \times 13.595 = 1033.22$ grams. The acceleration of gravity (normal) in c.g.s. units is 980.665. Multiplying the mass by the acceleration of gravity, we obtain for the pressure in c.g.s. units

$$1{,}013{,}250 \text{ dynes/sq. cm.}$$

Since dynes per square centimeter is an exceedingly small unit, V. Bjerknes introduced the *millibar*, which is 1000 times greater than the c.g.s.-unit. Thus, to a pressure of 76 cm. mercury corresponds 1013 mb. Pressure expressed in length units may be converted to millibars by the aid of Table I (Appendix).

Variation in Pressure with Elevation.—The difference in pressure between two points, the one above the other, is simply equal to the weight of the air column between the two points. Let Δp denote the measured difference in pressure, and let Δz denote the difference in altitude. Then

$$(1) \qquad -\Delta p = \rho g \,\Delta z$$

when ρ is the density of the air and g is the acceleration of gravity. The minus sign indicates that the pressure decreases as the altitude increases. Since the density of the air depends on temperature and pressure, we may tabulate the decrease in pressure per unit length along the vertical as a function of pressure and temperature. Such values are given in Table II (Appendix). Near the earth's surface, it is sufficiently accurate to say that the pressure decreases 4 mb./100 ft. ascent.

Equation (1) is sufficiently accurate if the difference in elevation is relatively small. For greater differences the following formula will hold:

$$\Delta z = (49{,}080 + 107t)\,\frac{p_0 - p}{p_0 + p} \tag{2}$$

Here Δz is the difference in altitude expressed in feet, t the mean temperature (degrees Fahrenheit) of the air column, p_0 the pressure at the lower end, and p the pressure at the upper end of the air column. The pressure variation along the vertical is about 10,000 times as great as in the horizontal direction. It is, therefore, not important that the observations be made along a vertical

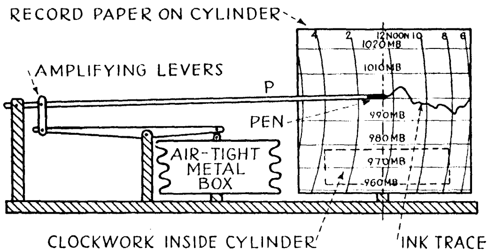


FIG. 5.—Showing the principle of the aneroid barograph. If the rotating cylinder were replaced by a pressure scale, the instrument would be an aneroid barometer.

line. Even when an aircraft is at a considerable distance from the airport, the pressure and temperature measured in the aircraft and at the airport will give sufficient accuracy for the evaluation of the height.

If the temperature were 50°F. (10°C.) throughout the air column, the above formula reduces to

$$\Delta z = 54{,}430\,\frac{p_0 - p}{p_0 + p} \tag{3}$$

which in most cases is sufficiently accurate.

The Aneroid Barometer.—For many purposes, it is inconvenient to use a mercurial barometer. A cheaper and more robust instrument is the aneroid barometer, where the atmospheric pressure is balanced by elasticity forces. An airtight metal box, from which air has been partly evacuated, is fixed to a frame (see Fig. 5). When the atmospheric pressure increases the box is slightly compressed, and when the pressure decreases the box expands. A pointer P connected with the upper face of the box and pointing towards a scale will then indicate the atmospheric pressure.

The Barograph.—A self-recording barometer is called a barograph. The commonly used barographs are of the aneroid type. In Fig. 5, the scale is shown on a record paper which surrounds a cylinder. Inside the cylinder is a clock that rotates the cylinder once a day (or once in 7 days). If an ink pen is placed on the end

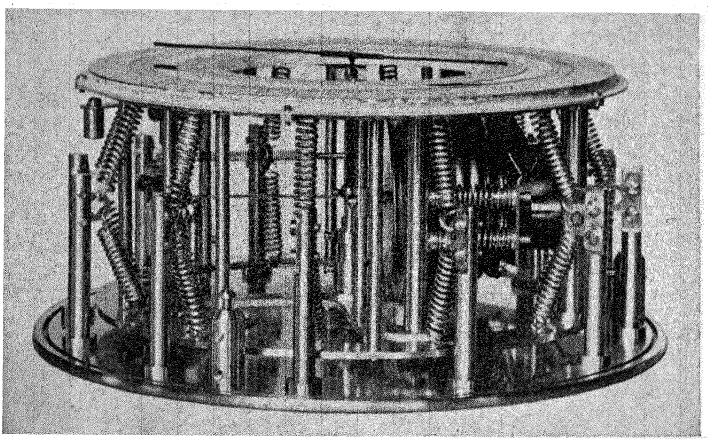


FIG. 6.—Showing the inside of an aneroid barometer. The springs eliminate vibrations. (*Courtesy of J. P. Friez & Sons, Inc.*)

of the pointer P, then a continuous curve is produced that shows the pressure at any given instant.

The Altimeter.—On account of the relation between pressure and altitude [Eq. (3), page 13], the aneroid barometer may be graduated to show altitude instead of pressure. The instrument is then called an altimeter.

It should be borne in mind that the altimeter is a *relative* and not an *absolute* instrument. Suppose, for example, that an aircraft is at sea level, where the pressure is 1010 mb. and the altimeter shows zero. If the pressure at sea level falls from 1010 to 1002 mb., the altimeter would indicate an altitude of about 200 ft. If the pressure again increased to, say, 1022 mb., the altimeter would indicate an altitude of about minus 300 ft. Thus, the altimeter indicates correct altitude only when it is set for the correct pressure at the surface.

Let us suppose that an aircraft starts from New York with correct altimeter reading and that the pressure in New York reduced to sea level is 990 mb. Suppose next that the aircraft flies to Chicago where the pressure reduced to sea level is 1030 mb. On arrival in Chicago the altimeter reading would be approxi-

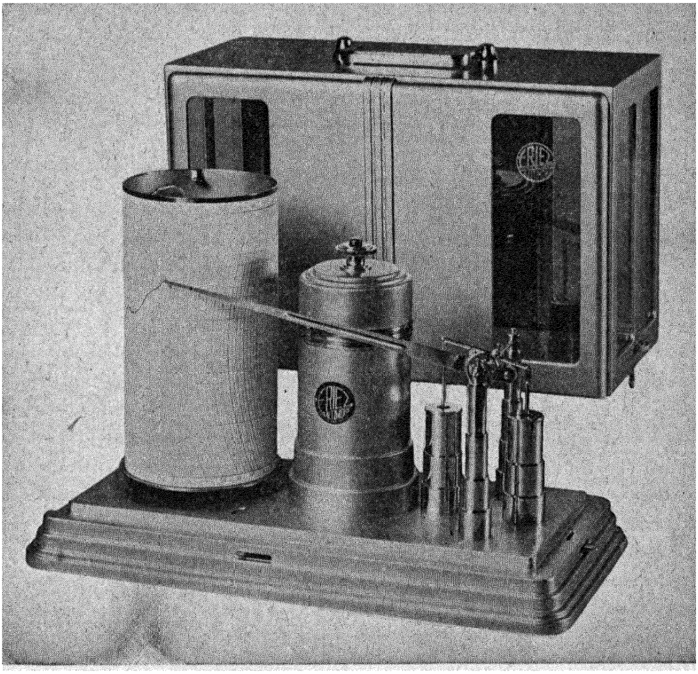


FIG. 7.—The barograph. (*Courtesy of J. P. Friez & Sons, Inc.*)

mately 1000 ft. too low, as can be seen from Table II (Appendix). The altimeter correction is due to the fact that falling pressure has the same effect on the altimeter as if the aircraft ascended and rising pressure the same effect as if the aircraft descended through the atmosphere.

Temperature.—The temperature of the air is usually measured by means of a mercurial thermometer. On land stations the instrument is hung in a louvered wooden screen in order to provide ventilation and to protect the instrument from the influences of radiation and precipitation. Special thermometers are constructed to record the maximum and the minimum temperatures.

The thermometers are usually graduated on the centigrade (C.) or the Fahrenheit (F.) scale. On the Fahrenheit scale the freezing point of pure water is at 32 and the boiling point at 212°. On the centigrade scale the freezing point is at 0 and the boiling point is at 100°. To convert Fahrenheit degrees to centigrade degrees, subtract 32 and multiply by $5/9$, thus:

$$C. = (F. - 32)5/9 \qquad \text{or} \qquad F. = 32 + 9/5C.$$

Table III (Appendix) gives the corresponding values of the two scales.

There is also a third scale in use, the so-called *absolute* scale (A.), where the freezing point is at 273, and the boiling point at 373°. The absolute temperature differs from the centigrade temperature by the constant 273. The advantage of the absolute scale is that negative temperatures cannot occur and that for any given pressure the density of the air is inversely proportional to the absolute temperature.

The centigrade scale is in general use in most European countries except Great Britain. The English-speaking countries use the Fahrenheit scale for reporting temperatures near the earth's surface and the centigrade scale for reporting temperatures in the free atmosphere. The absolute scale is mostly used in scientific literature.

The Thermograph.—A self-recording thermometer is called a thermograph. Most widely used is the bimetallic thermograph, the principle of which is illustrated in Fig. 8. An element consisting of two curved sheets of metal of widely different thermal expansion is fixed to a frame. When the temperature varies, the curvature of the bimetallic strip will change, and the free end will move up or down. This motion is transferred by amplifying levers to a pointer which is hinged to the frame. The pointer is furnished with an ink pen that writes on the temperature chart fixed to the cylinder which is kept in rotating motion by means of a clock.

Humidity.—The maximum amount of water vapor that the air can absorb depends entirely on the temperature; the higher the temperature of the air, the more water vapor can it hold. The water vapor, being a gas, contributes to the atmospheric pressure by an amount e that is called the partial pressure of the water vapor. The air is saturated with moisture when the partial pressure of the water vapor has reached the maximum amount that is possible at any given temperature. This maximum vapor pressure is called the saturation vapor pressure E.

Figure 9 shows diagrammatically the relation between the air temperature and the saturation vapor pressure. Suppose that we have observed a certain air temperature and a certain partial pressure of the water vapor. Plotting this in Fig. 9, we obtain a point D, which represents the physical conditions of the air. The

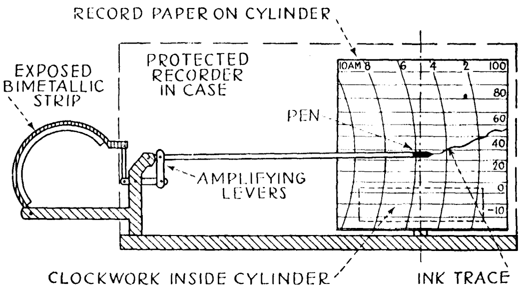


FIG. 8.—Showing the principle of the thermograph.

vertical distance from A to D shows the actual water vapor pressure e, and the vertical distance from A to B shows the saturation vapor pressure E at the temperature T. In other words, the vertical distance from D to B shows the *saturation deficit*, or how far the air is from saturation.

The *relative humidity* is defined as the ratio (expressed as a percentage) of the moisture content e of the air to the amount of moisture the air could contain E if it were saturated at the same temperature. Let R denote the relative humidity. Then, from Fig. 9,

$$R = 100 \frac{e}{E} \text{ (per cent)}$$

We consider again the point D in Fig. 9. Suppose we cool the air with constant moisture content. The point that represents the physical condition of the air will then move from D toward C. At C the air is saturated with moisture, and its temperature T_d is called the *dew-point temperature.* Thus, the dew-point temperature is the temperature to which the air must be cooled with constant moisture content in order to become saturated. While the air is cooled from D to C, the dew-point temperature stays constant. However, if the air is cooled below its dew point, condensation occurs, water vapor is removed from the air, and the dew point decreases. The dew-point temperature is, therefore, conservative with respect to temperature variation as long as the air is nonsaturated. The closer the dew point is to the actual temperature, the greater is the likelihood of formation of fog or clouds.

The moisture content of the air can also be expressed in terms of weight. The *absolute humidity* a is defined as the number of

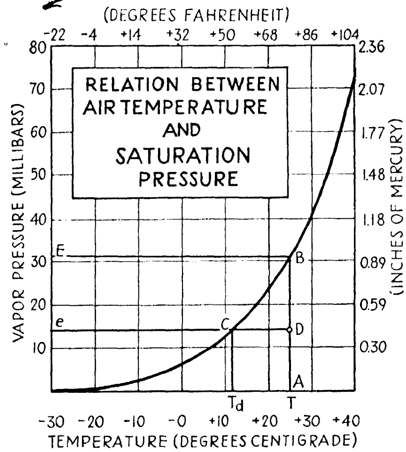


FIG. 9.—Showing the definition of relative humidity and dew-point temperature.

grams of water vapor contained in 1 cu. m. of natural air. It may be computed from the following approximate formula:

$$a = 217 \frac{e}{T} \text{ grams/cu. m.}$$

where e is the water vapor pressure and T the absolute temperature. It will be seen that absolute humidity varies both with the temperature and with the vapor pressure. It is, therefore, not conservative with respect to temperature changes and is seldom used in meteorology.

The *specific humidity* is defined as the number of grams of water vapor contained in 1 kg. of natural air. The specific humidity q can be computed from the atmospheric pressure p and the water vapor pressure e by means of the following formula:

$$q = \frac{623e}{p - 0.377e}$$

Finally, the moisture content of the air may be expressed as the number of grams of water vapor admixed with 1 kg. of dry air. This is called the *mixing ratio* and is computed from the formula

$$m = \frac{623e}{p - e}$$

The specific humidity differs so slightly from the mixing ratio that the difference between them is insignificant in view of the inaccuracy in the humidity measurements.

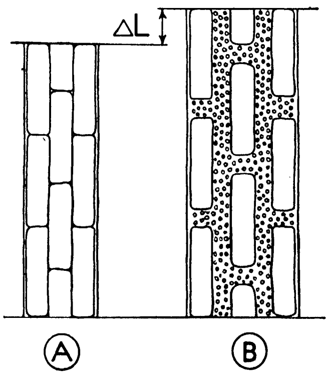


Fig. 10.—The length of a human hair increases with the relative humidity.

Humidity Instruments.—The moisture content of the air is obtained either from a hair hygrometer or from the readings of a "dry-bulb" and a "wet-bulb" thermometer.

The length of a human hair varies with the moisture in the air, as shown in Fig. 10. When the hair is dry (Fig. 10*A*) the individual cells are close together, but when the hair is moist (Fig. 10*B*) the space between the cells adsorbs water vapor and the hair lengthens by the amount ΔL. The *hair hygrometer* is an instrument for measuring the variations in the length of a bundle of hairs; it is graduated to indicate relative humidity instead of the length of the hair. A self-recording hygrometer is called a *hygrograph,* and its principle is illustrated in Fig. 11. The bundle of hairs is fixed to the frame at one end where there is also an adjusting screw that allows the observer to reset the zero point. The other end of the hairs is connected by amplifying levers to compensating cams which are, in turn, connected to a pointer and an ink pen which records variations on a paper chart. The chart, or record paper, surrounds the same type of recording apparatus as in the barograph and thermograph.

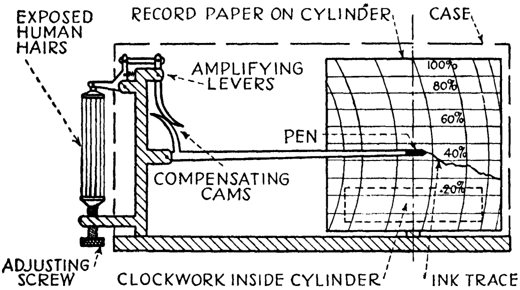


FIG. 11.—Showing the principle of the hair hygrograph. The compensating cams make the humidity scale linear.

A wet-bulb thermometer is an ordinary thermometer whose bulb is kept wet by a piece of muslin and a wick that dips into a

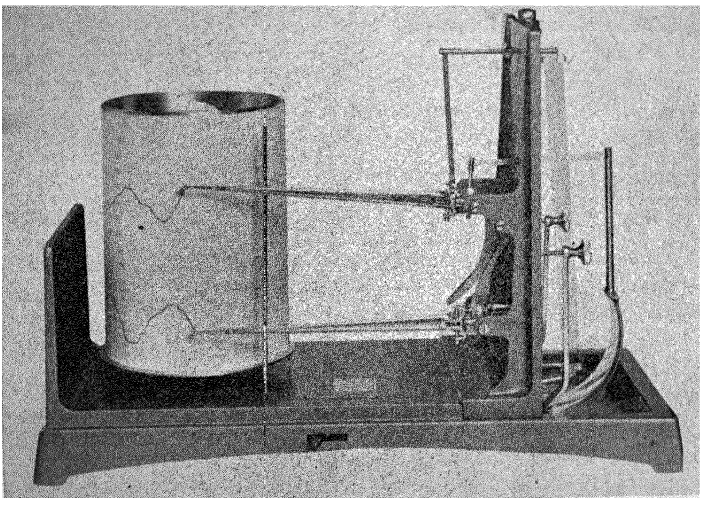


FIG. 12.—The thermohygrograph records temperature and relative humidity. (*Courtesy of J. P. Friez & Sons, Inc.*)

vessel containing pure water. If the air is not saturated with moisture, water will evaporate from the bulb of the wet thermometer. Since energy is required to evaporate water, the evaporation cools the wet thermometer so that it shows a lower temperature than that of the dry thermometer. The drier the air, the more intense is the evaporation, and the larger is the difference between the temperatures of the dry and the wet thermometer. From the readings of the two thermometers, the moisture content is readily obtained by means of humidity tables.

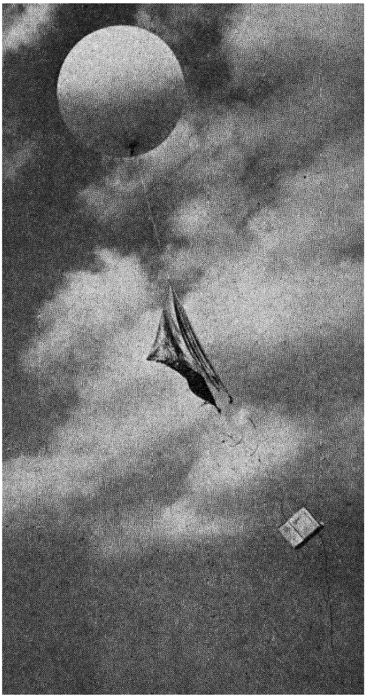


Fig. 13.—Balloon with radio-meteorograph for measuring pressure, temperature, and humidity in the free atmosphere. (*Courtesy of J. P. Friez & Sons, Inc.*)

Meteorographs.—In principle, the meteorograph consists of a barograph, a thermograph, and a hygrograph combined into one instrument, which produces three curves showing how pressure, temperature, and humidity vary with time. Such instruments are used by airplanes for measuring the conditions in the free atmosphere. On account of the relations between pressure and temperature on the one hand, and elevation on the other (see page

12), the records can be evaluated to show how temperature and humidity vary with elevation in the free atmosphere.

Airplanes carrying meteorographs do not normally reach beyond 20,000 ft. above sea level. In order to observe the conditions in the upper atmosphere, a new type of meteorograph—the radiometeorograph, or the radio sonde—has been constructed. This instrument has no clock and drum, and the instrument does not record mechanically. Instead, it is furnished with a minute radio sender which transmits to the ground radio signals indicating the values of pressure, temperature, and humidity. These instruments are carried by large balloons (see Fig. 13) which

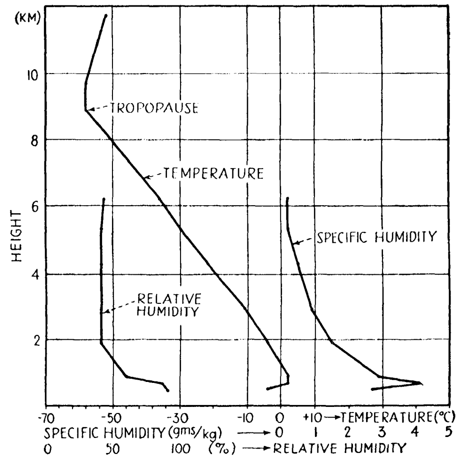


FIG. 14.—Distribution of temperature, relative humidity, and specific humidity as obtained from a radiometeorograph.

normally reach to altitudes of 40,000 ft. before they burst. Occasionally such balloons have reached altitudes exceeding 100,000 ft.

Figure 14 shows, as an example, the distribution of temperature and humidity along the vertical as recorded by a radiometeorograph. Note the change in the temperature curve as the balloon ascends through the tropopause.

Wind Direction and Velocity.—The wind direction is the direction from which the wind blows, a south wind being a wind blowing from true south. The wind directions observed at all land and ship stations are referred to the compass and expressed by letters or numerals, as shown in Fig. 15*A*. The wind directions in the free atmosphere are expressed in decades of degrees, as shown in Fig. 15*B*.

The wind direction (at land stations) is indicated by a freely exposed wind vane, which usually is connected with a recording apparatus indoors. The speed of the wind is recorded by an anemograph. Most commonly used is the cup anemometer shown in Fig. 16. The stronger the wind, the faster is the rotation of the cups. For each mile the wind has blown, an electric contact is made and transmitted to the recording instrument.

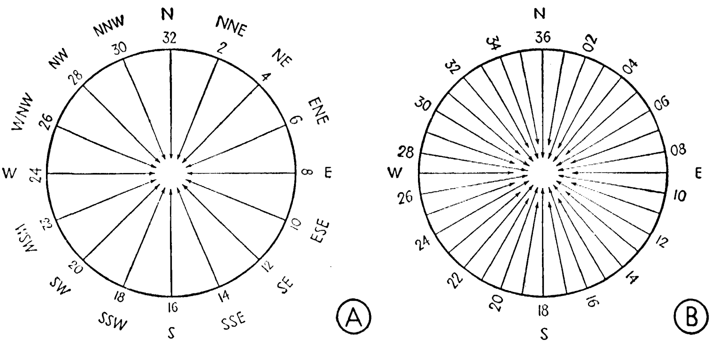


Fig. 15.—Wind directions. *A*, in surface reports; *B*, in reports from the free atmosphere.

The speed of the wind is expressed in meters per second, kilometers per hour, miles per hour, or knots. The relation between these units is

$$1 \text{ m./sec.} = 3.6 \text{ km./hr.} = 2.24 \text{ m.p.h.} = 1.94 \text{ knots}$$

Wind velocities expressed in one unit may be converted to other units by means of Table IV (Appendix).

It is possible to estimate the speed of the wind without the use of any instrument. The skilled observer will be able to estimate the force of the wind from the action that it has on certain objects, such as smoke, flags, and sails. When this is done, the wind speed is referred to what is called the Beaufort scale. This scale, which was introduced by Admiral Beaufort in 1805, is still in international use, and wind velocities that are measured are converted to Beaufort numbers in order to have a uniform standard throughout the *réseau* (or network) of stations.

The wind velocity varies with distance above the ground, and the variation is particularly rapid close to the ground. Table I gives the Beaufort scale of wind force, with velocity equivalents at about 6 m. (20 ft.) above level ground.

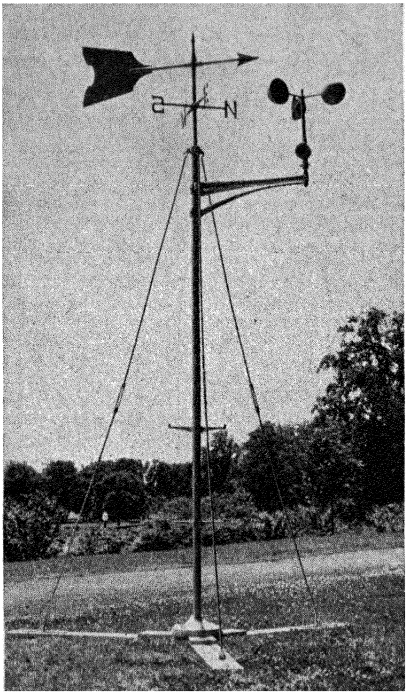


FIG. 16.—Cup anemometer and wind vane. (*Courtesy of J. P. Friez & Sons, Inc.*)

The wind is not a steady current. The velocity varies in a succession of gusts and lulls of variable direction. These variations of short periods are primarily caused by the roughness of the earth's surface which creates eddies that travel with the general current, superimposed on it. Larger irregularities in the wind are caused by convectional currents in unstable air masses. Such currents will be discussed in later sections.

The turbulence of the wind is an important agent in carrying water vapor, heat, dust, etc., up to high levels.

TABLE I.—THE BEAUFORT SCALE OF WIND FORCE WITH SPECIFICATIONS AND VELOCITY EQUIVALENTS

Beaufort number	General description	Specifications	Limits of velocity 6 m. above level ground			
			m./sec.	km./hr.	m.p.h.	Knots
0	Calm	Smoke rises vertically	0.3	Less than 1	Less than 1	Less than 1
1	Light air	Wind direction shown by smoke drift but not by vanes	0.6– 1.7	2– 6	1– 3	1– 3
2	Slight breeze	Wind felt on face; leaves rustle; ordinary vane moved by wind	1.8– 3.3	7– 12	4– 7	4– 6
3	Gentle breeze	Leaves and twigs in constant motion; wind extends light flag	3.4– 5.2	13– 18	8–11	7–10
4	Moderate breeze	Dust and loose paper; small branches are moved	5.3– 7.4	19– 26	12–16	10–14
5	Fresh breeze	Small trees in leaf begin to sway	7.5– 9.8	27– 35	17–22	15–19
6	Strong breeze	Large branches in motion; whistling in telegraph wires	9.9–12.4	36– 44	23–27	19–24
7	Moderate gale	Whole trees in motion	12.5–15.2	45– 55	28–34	24–30
8	Fresh gale	Twigs broken off trees; progress generally impeded	15.3–18.2	56– 66	35–41	30–35
9	Strong gale	Slight structural damage occurs; chimney pots removed	18.3–21.5	67– 77	42–48	36–42
10	Whole gale	Trees uprooted; considerable structural damage	21.6–25.4	78–90	49–56	42–49
11	Storm	Very rarely experienced; widespread damage	25.5–29.0	91–104	57–67	49–56
12	Hurricane	. . . .	Above 29.1	Above 105	Above 68	Above 56

Pilot-balloon Observations.—The winds in the free atmosphere are measured by means of small balloons filled with hydrogen or helium. These balloons are so made that they ascend with a constant velocity. Suppose that the ascensional velocity of the balloon is 600 ft./min. and that the velocity of the wind is V. In 1 min. the balloon will have ascended the vertical distance $PB = 600$ ft. and drifted with the wind the horizontal distance OP (see Fig. 17). A theodolite at O measures the angle every minute. By simple trigonometry,

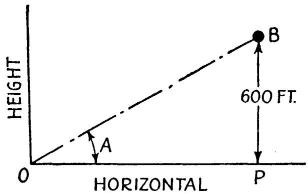


Fig. 17.

$$OP = V = 600 \text{ cotan } A \text{ (feet per minute)}$$

The theodolite also measures the azimuth angle from, say, north to the direction from O to P, and the direction of the wind is found. By observing the azimuth and the vertical angle every minute or every half minute and plotting the path of the balloon on a chart, the direction and the speed of the mean wind are found for any level.

Table II.—Classification of Clouds

Name	Abbreviation	Altitude
Cirrus	*Ci*	High
Cirro-stratus	*Cs*	
Cirro-cumulus	*Cc*	
Alto-stratus	*As*	Medium
Alto-cumulus	*Ac*	
Strato-cumulus	*Sc*	Low
Nimbo-stratus	*Ns*	
Stratus	*St*	
Cumulus	*Cu*	Vertical development
Cumulo-nimbus	*Cb*	

Classification of Clouds.—Even though the number of forms that clouds may take is almost unlimited, it is a fact that the number of *types* of cloud is limited. The international classification of clouds consists of 10 main types, which for convenience are arranged according to height above the ground in the manner shown in the table on page 26.

Cirrus (*Ci*) is the highest of all clouds. It has a typical fibrous (threadlike) structure and a delicate silky appearance (Fig. 18). Cirrus clouds are sometimes arranged irregularly in the sky as detached clouds without connection with cirro-stratus or alto-

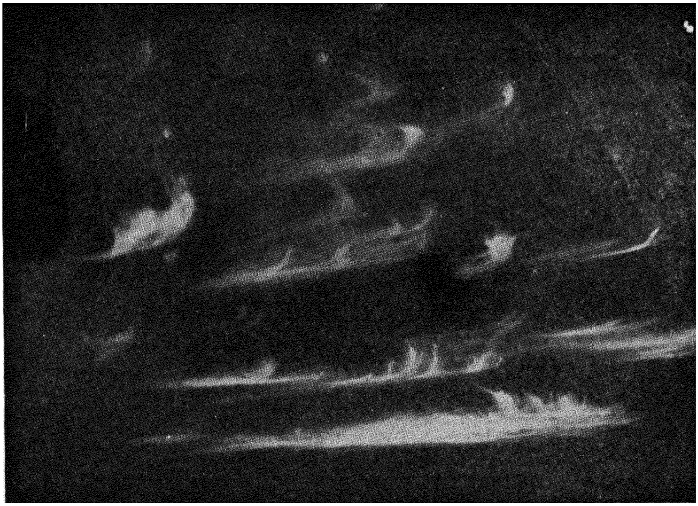


FIG. 18.—Cirrus, white and silky streaks. (*International Atlas of Clouds.*)

stratus. They are then called fair-weather cirri. If the cirrus clouds are arranged in bands or connected with cirro-stratus or alto-stratus or otherwise systematically arranged, they are usually the harbinger of bad weather. In "thundery" or squally weather, a special kind of cirrus (cirrus densus) is frequently observed which originates from the anvils of cumulo-nimbus. These clouds are often called *false cirrus*, because they are denser and usually lower than the ordinary cirrus.

Cirro-stratus (*Cs*) is a thin, whitish sheet of cloud, sometimes like a veil covering the whole sky and merely giving it a milky appearance, at other times showing signs of a fibrous structure like a tangled web (Fig. 19). Cirro-stratus often produces a halo around the sun or moon. It is often a sign of approaching bad weather.

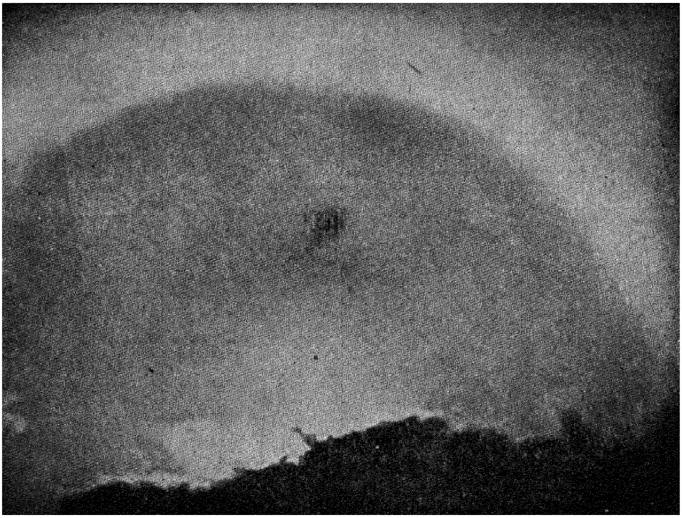


FIG. 19.—Cirro-stratus, thin, white, silky veil. Below, tops of cumulus. (*International Atlas of Clouds.*)

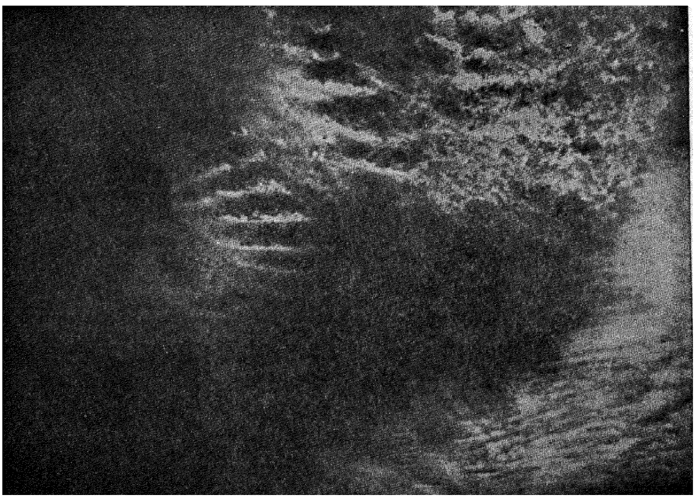


FIG. 20.—Cirro-cumulus, or mackerel sky. (*International Atlas of Clouds.*)

Cirro-cumulus (*Cc*) consists usually of small, white flakes of clouds without shadow, arranged in a regular pattern (Fig. 20). Cirro-cumulus develops from cirro-stratus. The pattern is due to a single or a double undulation of the cloud sheet.

Alto-stratus (*As*) is a dense sheet of gray or bluish color, frequently showing a fibrous structure (Fig. 21). It often merges gradually into cirro-stratus. Increasing alto-stratus is usually followed by precipitation of a continuous and lasting type.

FIG. 21.—Alto-stratus, often with a bluish-gray color. (*International Atlas of Clouds.*)

Alto-cumulus (*Ac*) differs from cirro-cumulus in consisting of larger globules, often with shadows, whereas cirro-cumulus clouds show only indications of shadows or none at all (Fig. 22). Alto-cumulus often develops from dissolving alto-stratus. An important variety of alto-cumulus is called *alto-cumulus castellatus.* In appearance, it resembles ordinary *Ac*; but, in places, turreted tops develop that look like miniature cumulus. Alto-cumulus castellatus usually indicates a change to a chaotic, thundery sky.

Strato-cumulus (*Sc*) is a cloud layer consisting of large lumpy masses or rolls of dull gray color with brighter interstices (Fig. 23). The masses are often arranged in a regular way and resemble alto-cumulus.

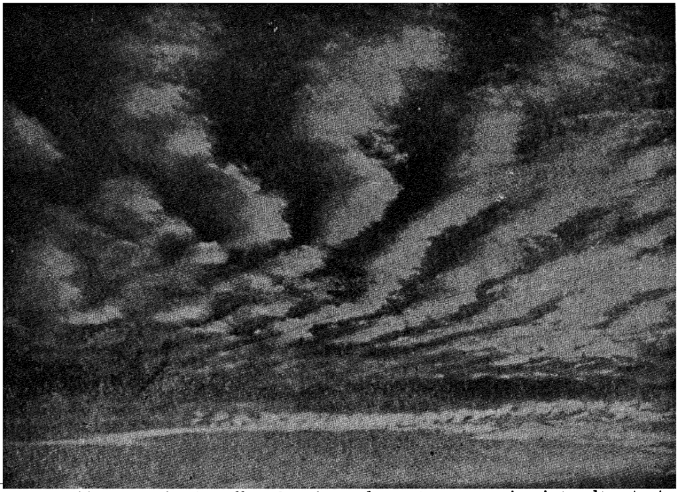


FIG. 22.—Alto-cumulus in rolls, often irregular pattern merging into alto-stratus. (*International Atlas of Clouds.*)

FIG. 23.—Strato-cumulus resembles alto-cumulus but is lower and usually grayer. (*International Atlas of Clouds.*)

Nimbo-stratus (*Ns*) is a dense, shapeless, and ragged layer of low clouds from which steady precipitation usually falls. It is usually connected with alto-stratus that is present above the nimbus. Fragments of nimbus that drift under the rain clouds are called *fractonimbus* or *scud.*

Cumulus (*Cu*) is a thick cloud whose upper surface is dome-shaped, often of a cauliflower structure, the base being usually horizontal. Cumulus clouds may be divided into two classes. Flat cumulus clouds without towers or protuberances are called

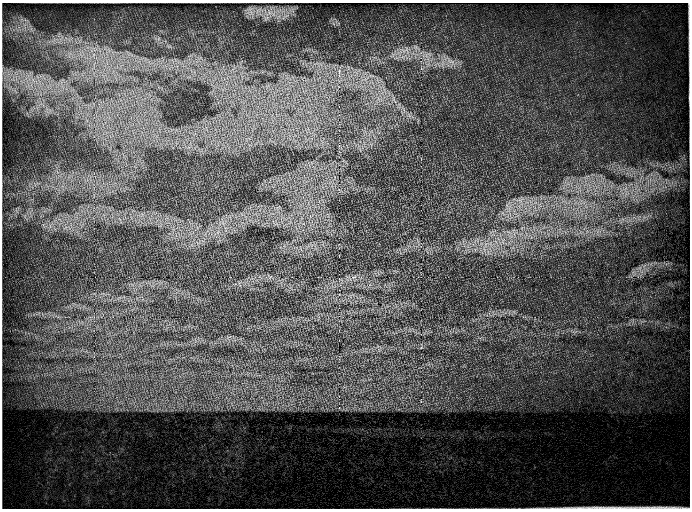


FIG. 24.—Cumulus humilis, or fair-weather cumulus, have no towers or protuberances. (*International Atlas of Clouds.*)

cumulus humilis or fair-weather cumulus (Fig. 24). Towering cumulus clouds with typical cauliflower structure showing internal motion and turbulence are called *cumulus congestus* (Fig. 25). They may develop into cumulo-nimbus.

Cumulo-nimbus (*Cb*) thunderclouds or shower clouds are great masses of cloud rising like mountains, towers, or anvils and having a base that looks like a ragged mass of nimbo-stratus. The tops are often anvil-shaped or surrounded by false cirrus. Figure 26 shows a cumulo-nimbus without anvil, and Fig. 27 shows one with anvil. The cumulo-nimbus clouds are accompanied by showers, squalls, or thunderstorms and sometimes hail. The *line squall*

FIG. 25.—Cumulus congestus, or towering cumulus; cauliflower structure. (*International Atlas of Clouds.*)

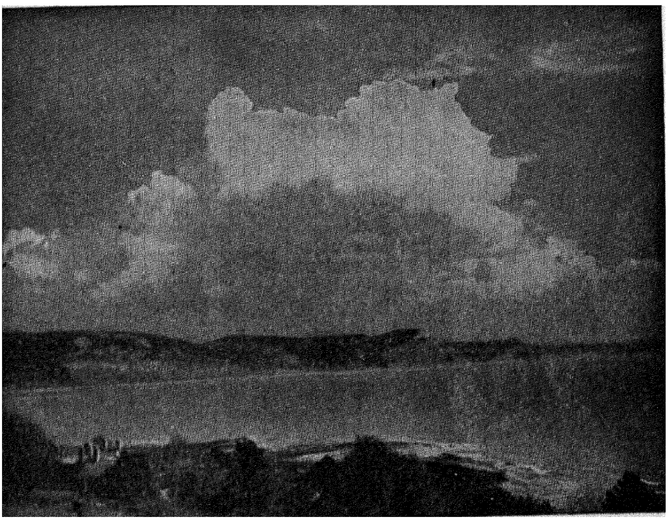


FIG. 26.—Cumulo-nimbus without anvil (calvus). (*International Atlas of Clouds.*)

cloud is a variety of cumulo-nimbus that extends like a long line or arch across the sky (Fig. 28).

Stratus (*St*) is a uniform layer of low foglike cloud, but not lying on the ground (Fig. 29). Seen from above, the stratus layers often show surges like a large swell on an ocean of cloud (see Fig. 30).

The heights of the various types of cloud vary within wide limits. The high clouds are usually above 20,000 ft. (6000 m.) and below 30,000 ft. (10,000 m.). The medium clouds are most frequently between 8000 and 20,000 ft. (2500 and 6000 m.), and the low clouds below 8000 ft. (2500 m.). The tops of cumulus

FIG. 27.—Cumulo-nimbus with anvil (incus). (*International Atlas of Clouds.*)

clouds, notably those of cumulo-nimbus, may reach to great heights, whereas their bases on the average are at or below 3000 ft. (1000 m.) above the ground.

Fog and Mist.—Fog is formed when the air near the earth's surface is cooled below its dew point. Fog is, therefore, nothing but a cloud that touches the ground. By agreement, such a cloud is called *fog* only when the visibility is less than 1 km. (5/8 mile); if the visibility is greater than this, it is called *mist*.[1]

The following scale is used to indicate the density of the fog:

Description	Visibility
Dense fog	Less than 50 yd.
Thick fog	50 to 200 yd.
Medium fog	1/8–1/3 mile
Moderate fog	1/3–5/8 mile
Mist	Above 5/8 mile

[1] This definition corresponds to the British use and the one occurring in the publications of the International Meteorological Organization. In North America, the word mist is often used synonymously with drizzle or fine rain.

The processes that lead to the formation of fogs will be discussed later.

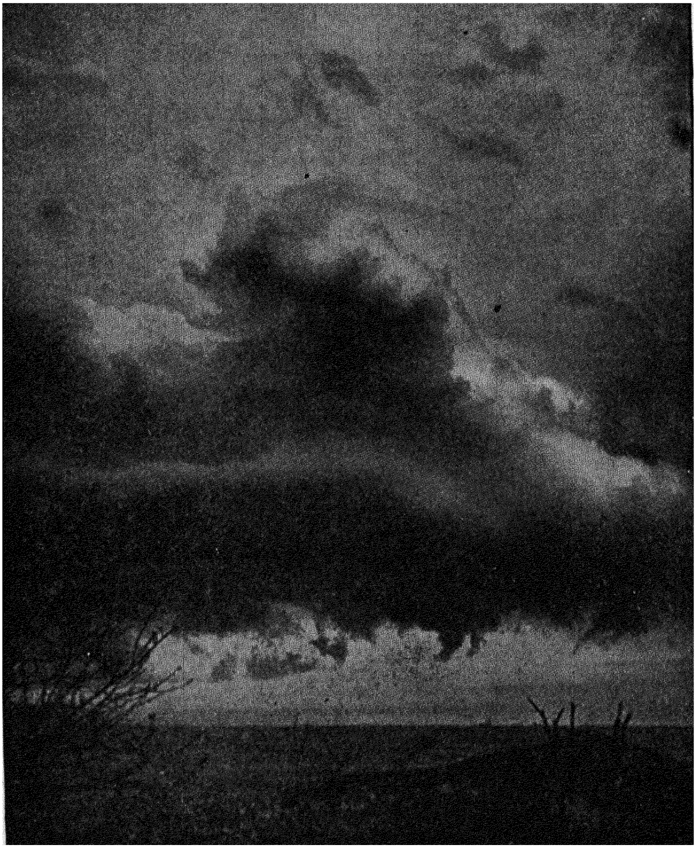


FIG. 28.—Cumulo-nimbus arcus, or the line squall cloud. (*International Atlas of Clouds.*)

Haze.—Haze consists of dust particles from the continents or of salt particles from the spray of oceans. These particles are too small to be seen individually by the naked eye, but their effect on visibility and on coloring of distant objects is easily observed. Through haze, distant objects are seen as if through a thin veil of pale blue if the object is dark and a yellowish veil if the object is white. At a certain distance, depending on the density of the

Fig. 29.—Stratus, low and gray cloud deck often touching the hills. (*International Atlas of Clouds.*)

Fig. 30.—Surges on a layer of stratus (or fog) indicating the topography beneath.

haze, all details of the landscape and all details of color disappear, and the objects stand out like a silhouette against the sky. In fair weather with sunshine on the object, the characteristic distance at which the details disappear is one-third of the distance at which the contours of a mountain can be distinguished; in dull overcast weather, the distance is only one-sixth of the total visibility.

Precipitation.—The immediate result of the condensation process is the formation of clouds or fogs consisting of numerous microscopically small water droplets. Under special conditions, which we shall describe later, several of these droplets (or ice crystals) coalesce and form large drops, snowflakes, or hailstones which fall through the air.

The various forms of cloud and precipitation reveal to the meteorologist valuable information as to the physical processes in the air. For this reason, the forms of precipitation are classified as follows:

1. *Drizzle* is rather uniform precipitation of very numerous minute drops (diameter less than 0.5 mm.), which seem almost to float in the air and visibly follow even slight air motion. Drizzle falls from fog or a thick layer of stratus.

2. *Rain* is precipitation of liquid water in which the drops, as a rule, are larger than in drizzle. Occasionally, the raindrops may be as small as the drizzle drops. This happens particularly when rain begins to fall from a high sheet of alto-stratus. The drops will then fall through a deep layer of unsaturated air, and they will tend to evaporate. However, in such cases, the small raindrops are not nearly so numerous as are the drizzle drops. This distinguishes drizzle from small raindrops. Moreover, small raindrops fall from a sheet of high clouds, whereas drizzle falls from low stratus or fog.

3. *Snow* is precipitation of solid water mainly in the form of branched hexagonal crystals or stars. At temperatures not far from freezing, they are usually matted together in large flakes.

4. *Sleet* (British) is precipitation consisting of melting snow or a mixture of snow and rain. Snow and sleet may melt completely while falling through the air. They will then appear as rain at the ground. On the other hand, rain that freezes when falling from warm air aloft into cold air at the ground will not appear as snow (see below).

5. *Grains of ice* (British), or *sleet* (United States), is precipitation of grains or pellets of ice. They are transparent and occur

when raindrops from warm air aloft fall through a layer of cold air near the ground. (Dangerous icing condition.)

6. *Glaze*, or *freezing rain*, is more or less transparent ice deposited on objects. The difference between grains of ice and glaze is that the former result from rain that freezes in the air and the latter from rain or drizzle that freezes after reaching the ground.

7. *Ice needles*, or diamond dust, are thin shafts or small sheets of ice which are so light that they seem to be suspended in the air like a frozen fog. When glittering in the sunshine, they become especially visible and often give rise to sun pillars or other optical phenomena. They occur only at very low temperatures.

8. *Granular snow* is precipitation of opaque small grains falling from stratus clouds. It is the frozen product that corresponds to drizzle.

Granular snow and ordinary dry snow do not form appreciable deposits on aircraft; but sleet, wet snow and subcooled waterdrops may form heavy deposits.

9. *Soft hail* consists of white, round grains of snowlike structure with diameter of 2 to about 5 mm. The grains are crisp, and they rebound and disintegrate easily.

10. *Small hail* consists of semitransparent, round or occasionally conical grains about 2 to 5 mm. in diameter. The grains generally consist of nuclei of soft hail surrounded by a thin crust of ice which gives them a glazed appearance. Small hail falls from cumulo-nimbus clouds, mostly at temperatures above freezing. It often falls together with rain and is, therefore, usually wet.

11. *Hail* is precipitation of balls or irregular lumps of ice of diameters ranging from 5 to 50 mm. or more. They are either transparent or composed of clear layers of ice alternating with opaque layers of snowlike structure. Real hail falls almost exclusively in violent thunderstorms and is very rare at temperatures below freezing at the earth's surface.

12. *Thunder and Lightning.*—When the difference in time is less than 10 sec., these phenomena are reported as occurring "at the station"; if the time difference is greater than 10 sec., they are reported as occurring in "the neighborhood of the station." When there is no audible thunder, the phenomenon is reported as "distant lightning."

13. *Dust-* or *sandstorms* consist of dust or sand raised by the wind to such an extent that the horizontal visibility is consider-

ably diminished. The dust and sand are never carried far from the source, and the conditions favorable for the formation of dust- and sandstorms are extreme dryness of the ground, unstable and turbulent air, and high wind velocity.

14. *Drifting snow* is reported when there is no real precipitation but snow from the ground is carried up into the air to such an

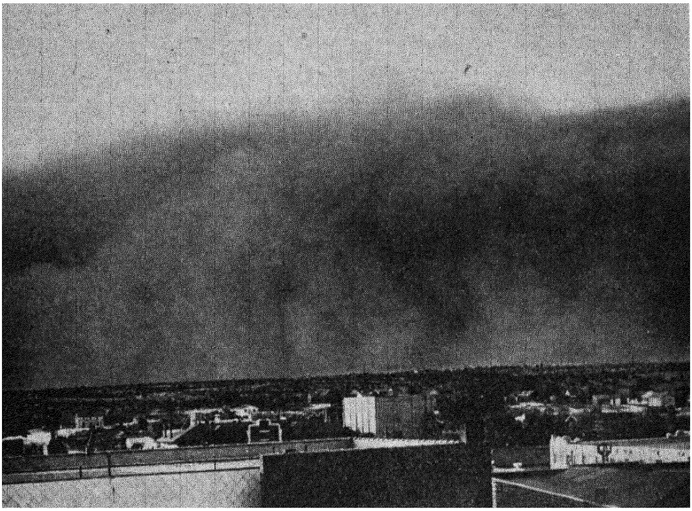


FIG. 31.—Approaching sandstorm, Sept. 14, 1930, Big Spring, Tex. (*Courtesy of U. S. Weather Bureau.*)

extent that the horizontal range of visibility is considerably diminished.

Clouds and Precipitation.—The above classification of precipitation is mainly based on the appearance of the precipitation elements without regard to the processes that lead to their formation. From the latter point of view, we may distinguish between three main groups, *viz.:*

1. *Intermittent or continuous precipitation* from a continuous cloud cover of the alto-stratus or nimbo-stratus type. This kind of precipitation is caused by the *slow* upward movement of a *large* mass of air, due to convergence in the horizontal motion of the air. For reasons that will be explained later, this type of precipitation is called *frontal precipitation.*

2. *Showers* (squalls, flurries), or precipitation of short duration that begins and ends suddenly, usually with fair periods between. This kind of precipitation, which originates from cumulo-nimbus clouds and is indicative of instability, is caused by the fairly *rapid* rising of *small* bodies of air through the atmosphere.

3. *Drizzle*, consisting of small elements falling from low cloud layers (stratus or fog). This kind of precipitation, which is indicative of stable air masses, is not connected with any appreciable ascensional velocity; on the contrary, the small drops fall out of the cloud because of the absence of any appreciable upward movement.

Clouds and precipitation will be discussed more fully in later sections on air masses, fronts, and cyclones.

Cloudiness.—A note is made at the hour of observation of how many tenths of the sky are covered with clouds. Owing to the importance of the low clouds for aviation, a special note is made of how many tenths of the sky are covered by low clouds.

Ceiling.—The height of the base of the low clouds above the ground is an important element, particularly for aviation. It may be measured by small ceiling balloons whose ascensional velocity is known. At night, it may be measured by attaching a light to the balloon or by the use of searchlights. Usually, the height of the clouds above the ground is estimated by the observer. In the international system of reporting, the following scale is used:

TABLE III.—HEIGHT OF THE BASE OF LOW CLOUDS ABOVE THE GROUND

Scale	Height, m.	Approximate equivalents, ft.
0	Less than 50	Less than 150
1	50– 100	150– 300
2	100– 200	300– 650
3	200– 300	650–1000
4	300– 600	1000–2000
5	600–1000	2000–3000
6	1000–1500	3000–5000
7	1500–2000	5000–6500
8	2000–2500	6500–8000
9	Greater than 2500	Greater than 8000

In the United States the ceiling is defined as the height in of the lowest level below 10,000 ft. above the ground at whicl total cloudiness covers more than one-half of the sky, except in the presence of heavy precipitation, dense fog, and other cc tions which prevent the observer from seeing any cloudiness may be present the ceiling is at zero altitude. Accordingly, ceiling is reported as "unlimited" when less than one-half ol sky below 10,000 ft. is covered by clouds.

Visibility.—The term visibility is used to describe the tı parency of the air in the horizontal direction. Visibility thus be defined as the greatest distance at which objects ca recognized, it being of course understood that they woul easily recognizable in a clear atmosphere.

The above definition refers to the horizontal daylight visibi In darkness, it is necessary to use lights of known candle pow determine the horizontal visibility.

The visibility is measured or estimated and reported on a s ranging from 0 to 9. Table IV gives the scale numbers. corresponding daylight visibility according to the above de tion, and also the corresponding distances at which a ligl 100 c.p. becomes indistinguishable.

TABLE IV.—VISIBILITY

Scale number	Daylight visibility	Night observations	
		Distance of object	Distance at wh light of 100 c becomes indistinguisha
0	Less than 50 m.	50 m.	100 m.
1	50– 200 m.	200 m.	330 m.
2	200– 500 m.	500 m.	740 m.
3	500–1000 m.	1 km.	1340 m.
4	1– 2 km.	2 km.	2.3 km.
5	2– 4 km.	4 km.	4.0 km.
6	4– 10 km.	10 km.	7.5 km.
7	10– 20 km.	20 km.	12 km.
8	20– 50 km.	At greater distances a 100-c.p. light i suitable	
9	Above 50 km.		

The visibility depends greatly on the weather. Table V shows the normal relation between visibility and some of the more frequent weather phenomena.

TABLE V.—NORMAL RELATION BETWEEN WEATHER AND VISIBILITY

Scale number	Daylight visibility	Fog, mist, or haze	Snow	Drizzle	Rain
0	Less than 50 m.	Dense	Very heavy		
1	50– 200 m.	Thick	Very heavy or heavy	....	Tropically heavy
2	200– 500 m.	Medium	Heavy	....	
3	500–1000 m.	Moderate	Moderate	Thick	Very heavy
4	1– 2 km.	Mist	Light	Moderate	Heavy
5	2– 4 km.	Slight mist or haze	Very light	Slight	Heavy
6	4– 10 km.	Slight mist or haze	Very light	....	Moderate
7	10– 20 km.	....	....	....	Light
8	20– 50 km.	....	....	....	Very light
9	Above 50 km.				

The haze is often arranged in layers in the atmosphere. In such cases the vertical visibility may vary greatly in different directions. A pilot flying in relatively pure air above a haze layer may not be able to see the ground, though the aircraft is perfectly visible from the ground. This condition is due to reflection and scattering of light from the top of the haze layer. In other cases, when there is a shallow layer of mist or fog at the ground, the pilot, flying above the layer, may see the ground directly under him but be unable to see the landing field when coming in to land. In such cases, the portion of the ground that is visible to the pilot increases with the altitude of the aircraft.

Use of Meteorological Observations.—It is essential that observations should be made simultaneously at all stations within large areas, so that they give an adequate picture of the state of the atmosphere at a given moment. Such observations, made at frequent intervals and plotted on weather maps, furnish the basis for the safeguarding of the airways as well as for general weather forecasting, storm warnings, and other forecasts. The weather map also furnishes an invaluable means for scientific research into the causes of the weather processes.

Observations are sometimes used to obtain mean values for shorter periods, say, 5 days, 10 days, or 1 month. Various maps of short-period mean values are now used as a basis for long-range forecasting.

A series of observations covering several years may be used to evaluate the *average* or *normal* values of temperature, pressure, rainfall, etc., and their periodical and aperiodical variations. A set of such normal values gives us what we call the *climate* of the station in question. Such normal values for a number of stations are used to construct maps that show the distribution of the various climatic zones over the earth's surface.

For aviation purposes, normal values are of but little interest, for in planning of new airways, flights over long distances, etc., it is far more important to know the frequency of visibility, ceiling, high winds, etc., than to know the climatological normals. Seasonal frequency tables for the more important elements have been published in many countries.

CHAPTER III

EVAPORATION, CONDENSATION, AND PRECIPITATION

Evaporation is the process that transforms water into water vapor. Condensation is the process that transforms water vapor into water. It is important to note that there is a principal difference between the condensation process and the precipitation process. Through condensation of water vapor, minute fog or cloud droplets, or dew, are formed. The precipitation process consists in the coalescence of several cloud or fog droplets into large drops. Therefore, condensation of water vapor does not necessarily lead to precipitation.

Evaporation.—To illustrate the process of evaporation, we consider a free water surface whose temperature is T. To any given temperature of the free water surface corresponds a saturation vapor pressure as shown by the saturation curve in Fig. 9. If the vapor pressure in the air above the water surface is less than the saturation value, then water will evaporate from the surface and mix into the air as water vapor. The evaporation process will continue until the partial pressure of the water vapor in the air equals the saturation vapor pressure at the water surface.

The speed with which water evaporates depends on (1) the difference between the saturation vapor pressure at the water and the actual vapor pressure in the air, (2) the wind velocity, and (3) the temperature. The stronger the wind and the higher the temperature, the more rapidly will the water evaporate.

The evaporation of water requires energy. Approximately 590 gram calories are required to bring 1 gram of the water from the liquid to the gaseous state. Conversely, when 1 gram of water vapor is condensed, the same amount of heat is liberated. This amount of heat is called the *latent heat of vaporization.*

Nuclei of Condensation.—Laboratory experiments have shown that condensation does not occur in perfectly pure air when the dew-point temperature is reached. The reason for this is that the cloud and fog droplets must have some sort of nuclei to form on. Air that is cooled below its dew-point temperature without condensation occurring is said to be *supersaturated* with moisture. In supersaturated air, the relative humidity exceeds 100 per cent. However, observations show that only minute degrees of supersaturation occur in the atmosphere.

The air contains a considerable amount of microscopically small dust particles, but most of these are useless as condensation nuclei. Only particles with an affinity to water vapor serve as condensation nuclei.

It is well known that calcium chloride strewn on sandy roads prevents dust from whirling up into the air. Calcium chloride has a great affinity to water, with the result that it absorbs water vapor from the air and moistens the roads even in dry weather. Any substance that has this affinity to water vapor is said to be *hygroscopic*. Only the hygroscopic particles in the air serve as nuclei of condensation.

Chemical analyses of cloud and fog water show that the condensation takes place on nuclei of sodium chloride (sea salt) and various sulphates. It is believed that minute particles of sea salt are brought into the air through evaporation of the spray of the oceans. The spray droplets evaporate, and the salt particles remain in the air.

Through the burning of fuels, sulphur dioxide is produced which oxidizes in the air and forms highly hygroscopic nuclei of sulphur trioxide. Thus, particles of sea salt and products of combustion containing sulphates are the main nuclei of condensation in the air.

The amount of hygroscopic material in a cloud droplet is, of course, very small. On the average, a cloud droplet contains 1 part hygroscopic material to 10,000 parts water by weight. An average cloud droplet is about 40 microns in diameter (1 micron equals 0.001 mm.), and an average condensation nucleus is about 1 to 2 microns. In comparison, it may be mentioned that the diameter of the raindrops varies from about 500 to about 4000 microns (*i.e.*, from 0.5 to 4 mm.).

The Condensation Process.—From a meteorological point of view, the essential features of the condensation process are (1) the absorption of water vapor by the hygroscopic nuclei and (2) the liberation of the latent heat of vaporization.

When the air is cooled so that it approaches the dew point, the hygroscopic nuclei begin to absorb water from the air. The larger nuclei will then cause condensation to occur even before the saturation point is reached. However, as the drops grow in size, they become so diluted that they become less and less active as hygroscopic material. The condensation process can then proceed only when the air is cooled slightly below its dew point so that a slight amount of supersaturation is present.

The growth of the fog and cloud droplets has been computed by Houghton from the equation of diffusion. The computations show that the small drops grow much more rapidly than the large drops. There is therefore a tendency for all drops to attain about the same size. When the drops have become relatively large, the rate of growth is exceedingly small. Thus, it would take a condensation nucleus only 100 sec. to grow to an average cloud droplet, but it would take a cloud droplet about 24 hr. to grow to the size of an average raindrop. Since a raindrop cannot be kept afloat in the cloud for such long intervals of time, it follows that the raindrops do not form through continued condensation; they must form through coalescence of several cloud droplets.

Since condensation liberates the latent heat of vaporization, the process tends to increase the air temperature and make the air relatively drier. To maintain condensation, it is therefore necessary to cool the air so much that it is kept in a slightly supercooled state. In other words, the cooling must overcompensate the liberation of the latent heat of vaporization. Condensation is therefore almost invariably a result of cooling of air masses. The processes that cause cooling of air masses will be discussed in the following chapter.

For the later discussion of clouds and precipitation, it is of interest to note that the water droplets in a cloud do not necessarily freeze when the temperature falls below freezing. Water droplets are normally present in the clouds even at temperatures much below freezing. The water is then *supercooled*, and it solidifies readily when the drops are disturbed. At temperatures about 20 to 0°F. the clouds usually consist of ice particles or a mixture of water droplets and ice particles.

At very low temperatures, water may be transformed from vapor to ice without passing through the state of liquid water. In meteorology, this process is usually called *sublimation*, although the word has a different meaning in physical chemistry. It is probable that certain nuclei (sublimation nuclei) are required to initiate sublimation, but not much is known of their properties. The reverse process is well known, *viz.*, when ice is transformed directly into water vapor. This happens frequently to the snow on the ground in dry winter weather.

The Precipitation Process.—The next item to consider is the mechanism that causes the cloud droplets to coalesce into large raindrops. When there is no coalescence, the cloud is said to be *colloidally stable.* Coalescence of the droplets is therefore a result of *colloidal instability.* There are five factors that influence the colloidal stability of a cloud, *viz.:*

1. The electric charge of the droplets.
2. The size of the droplets.
3. The temperature of the droplets.
4. The motion of the droplets.
5. The presence, or nonpresence, of ice crystals in the cloud.

In discussing the mechanism that causes precipitation to be released from the clouds, we meet with two problems, *viz.:* (*a*) how to explain the coalescence of the droplets and (*b*) how to explain the sudden release of precipitation that is usually observed.

1. *Electric Charge.*—The cloud droplets usually carry an electric charge; and when neighboring drops have opposite charges, the electric forces of attraction would tend to cause coalescence. However, the charges are so small and the distances between the droplets are such that no appreciable coalescence results. Moreover, it is difficult to conceive of any mechanism that would rearrange the charge distribution so suddenly as to account for the sudden release of precipitation that is often observed.

2. *Size.*—The saturation vapor pressure varies slightly with the curvature of the water surface: the larger the drop, the lower the saturation vapor pressure. In a cloud consisting of large and small drops the vapor pressure of the air would adopt a mean value, and there would be supersaturation over the larger drops and saturation deficit over the smaller drops, with the result that the smaller drops would tend to evaporate and condense on the larger drops. Computations show that this effect is altogether negligible.

3. *Temperature.*—It will be seen from Fig. 9 that the saturation vapor pressure varies with the temperature. The variation is but slight when the temperature is low, and it increases rapidly with increasing temperature. In a turbulent cloud, colder drops com-

ing from above would be brought into proximity with warm drops coming from below. There would then be supersaturation at the colder droplets and subsaturation over the warm droplets, with the result that the colder drops would grow at the expense of the warmer drops. This effect is of noticeable magnitude when the air temperature is high (50°F. or more); but it is altogether negligible at temperatures at, or below, freezing. Since the clouds usually are at high levels where the temperature is low, the effect of varying temperature cannot account for the coalescence, except in tropical regions where the temperature is sufficiently high.

4. *Motion.*—It is possible that turbulent motion may lead to collisions between the cloud droplets. But this cannot be the

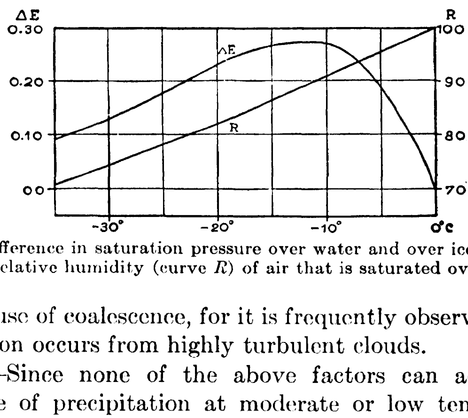


FIG. 32.—Difference in saturation pressure over water and over ice (curve ΔE); and relative humidity (curve R) of air that is saturated over ice.

general cause of coalescence, for it is frequently observed that no precipitation occurs from highly turbulent clouds.

5. *Ice.*—Since none of the above factors can account for the release of precipitation at moderate or low temperatures, Bergeron considered the colloidal instability that will result in a cloud consisting of a mixture of ice particles and subcooled water droplets.

The saturation curve shown in Fig. 9 refers to water. At temperatures below freezing, the saturation vapor pressure over ice (or snow) is lower than over subcooled water. This means that air that is saturated with respect to water is supersaturated with respect to ice. The difference between the saturation pressure over water and the saturation pressure over ice is shown by the curve ΔE in Fig. 32. In a cloud consisting of both water droplets and ice particles, the vapor pressure of the air would be a compromise between the two saturation pressures, with the result that the water droplets will evaporate and condense on the ice particles. Through this process, the ice elements become too heavy and fall through the cloud; in doing so, they collide with the water droplets and continue to grow until they leave the cloud.

Bergeron's theory for the release of precipitation is by far the most satisfactory one and is substantiated by a considerable amount of observational evidence. On the other hand, it has been observed that showers form at high temperatures without the cloud reaching up to subfreezing temperatures. These showers, which usually are quite light, can be accounted for as a result of turbulent mixing of warmer and colder drops in clouds whose temperature is sufficiently high.

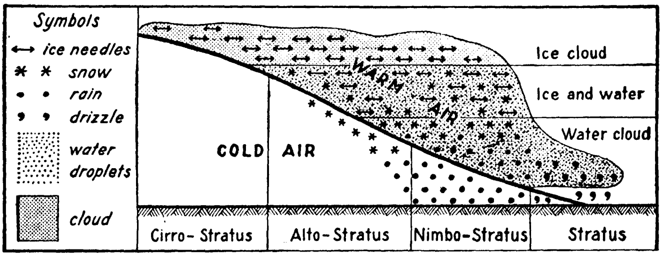


FIG. 33.—A rain cloud usually has ice particles in its top. (*After Bergeron.*)

Figure 33 shows diagrammatically the structure of a rain cloud. The upper portion of the cloud normally consists of ice crystals (cirrus clouds). Then comes a layer in which ice particles and water droplets are mixed. It is in this layer that the cloud is colloidally unstable. The elements that fall from this layer pick up water droplets as they fall through the water cloud below. If the cloud base is very high, the rain may evaporate before it reaches the ground.

The sudden release of precipitation may also be explained by Bergeron's theory. As long as the cloud consists of water droplets only, it is usually colloidally stable. However, when it builds up to such heights that ice crystals begin to form, it becomes colloidally unstable and precipitation is released. The ice particles have the same function in releasing precipitation as the condensation nuclei have in initiating condensation.

CHAPTER IV

ADIABATIC TEMPERATURE CHANGES

The processes that determine the weather are dependent partly upon the transfer of heat and moisture and partly upon the forces that create and maintain the motion in the atmosphere. In this chapter, we shall discuss such temperature changes as are due to expansion and contraction of air that moves upward or downward through the atmosphere. Air that moves upward comes under lower pressure and expands, whereas air that moves downward comes under higher pressure and contracts. Through contraction and expansion the temperature of the moving air changes even when no actual heat is supplied to or withdrawn from it. Such temperature changes are called *adiabatic* changes because no heat is added to, or withdrawn from, the air.

The Gas Law.—If in any given place the atmospheric pressure is p, the density ρ, and the absolute temperature T, the gas law may be expressed by the equation

$$p = \rho RT$$

where R is a numerical constant.

According to the molecular theory, a gas, or a mixture of gases, such as air, may be regarded as made up of a large number of minute molecules which are in a state of incessant and irregular motion leading to frequent collisions between the individual molecules. The effect of the impacts of the individual molecules produces the pressure of the gas. This pressure depends, therefore, on the number and the mass of the molecules and the speed with which they are moving. The number and the mass of the molecules define the density of the gas. It follows, then, that the pressure must be proportional to the density ρ. The average speed with which the molecules move depends on the absolute temperature. If the temperature were 0° absolute, the molecules would be at rest and there would be no pressure. The higher the temperature, the more the molecules are agitated and the greater is the frequency of the collisions. Therefore, the pressure must be proportional to the absolute temperature, as indicated in the above equation.

The quantity R, which is called the *gas constant*, characterizes the gas, or the mixture of gases, and its numerical value depends on the units used. If atmospheric pressure p is expressed in millibars and density ρ in metric tons per cubic meter, the gas constant for dry air is 2870.

Since the amount of water vapor in the atmosphere is variable, the gas constant varies a little with the moisture content; but this variation is slight and may be neglected unless great accuracy is required.

The First Law of Thermodynamics.—We consider a unit mass of air whose volume is α. Let c_v and T denote the specific heat and the absolute temperature, respectively. Suppose that the unit mass of air is within an airtight container, so that it can neither expand nor contract, and that a small amount of heat ΔQ is added to the air. Then

$$\Delta Q = c_v \, \Delta T$$

and all the added heat goes to increase the temperature of the air by the amount ΔT.

Since the temperature has increased by the amount ΔT, it follows from the gas law that the pressure within the container must have increased by the amount $\Delta p = R\rho \, \Delta T$.

Suppose, now, that the walls of the container are removed. The heated air will then expand, and its pressure will tend to adapt itself to the pressure of the surrounding air. In expanding, the volume increases by the amount $\Delta\alpha$, and the density decreases by the amount $-\Delta\rho$. In expanding, the air must do a certain amount of work against the external pressure, and this work is proportional to the external pressure p and the amount of expansion $\Delta\alpha$, or, in other words, it is equal to $p \, \Delta\alpha$. It follows, then, that part of the heat added to the air is used in the work incidental to the expansion, and only part of the added heat goes to increase the air temperature. Since no energy can be lost, it follows that *the added heat must be equal to the sum of the increase in internal energy and the energy consumed in the expansion.* Equating like unto like, we obtain:

$$\Delta Q = c_v \, \Delta T + p \, \Delta\alpha \tag{1}$$

or

Added heat = increase in internal energy + work due to expansion

This equation expresses the first law of thermodynamics.

Since α is the volume of the unit mass, it follows that $\alpha = 1/\rho$. The gas law may then be written as

$$\alpha p = RT \tag{2}$$

Furthermore, as was shown on page 12, the pressure variation along the vertical is expressed by

$$-\Delta p = \rho g \, \Delta z = \frac{g}{\alpha} \Delta z \tag{3}$$

This equation is called the equation of static equilibrium; the minus sign in front of Δp indicates that the pressure decreases by the amount Δp when the altitude increases by the amount Δz.

The first law of thermodynamics, the gas law, and the equation of equilibrium are the three laws that govern the relation between pressure, temperature, and density.

It may be noted that the equation of static equilibrium is not exact; it holds strictly only when the motion is not accelerated along the vertical, but the correction for vertical acceleration is altogether negligible.

Adiabatic Processes.—A process is said to be adiabatic when no heat is added to, or withdrawn from, the air that partakes in the process. In this case $\Delta Q = 0$, and therefore

$$-c_v \, \Delta T = p \, \Delta\alpha \tag{4}$$

This is the same as saying that the air cools while it expands and heats up when it contracts in an adiabatic process.

Near the earth's surface, the processes are mostly nonadiabatic, for the air receives heat from, or gives off heat to, the underlying surface. In the free atmosphere, the processes are mainly adiabatic, for the air is then far removed from any appreciable source of heat. Equation 4 is therefore of fundamental importance for the discussion of the temperature variations in the free atmosphere.

Ascent and Descent of Nonsaturated Air.—Suppose that a unit mass of air ascends a small distance Δz through the atmosphere and that no heat is added to, or withdrawn from, the air. As the air ascends, it comes under lower pressure and expands. Its pressure, temperature, and volume will then vary by the small amounts Δp, ΔT, and $\Delta\alpha$, respectively. Since the gas law must hold throughout the process, we may write

$$(\alpha + \Delta\alpha)(p + \Delta p) = R(T + \Delta T)$$

or

$$\alpha p + \Delta\alpha p + \alpha\,\Delta p + \Delta\alpha\,\Delta p = RT + R\,\Delta T$$

The product of the two small quantities $\Delta\alpha$ and Δp is very small compared with the other terms and may therefore be neglected. Furthermore, since

$$\alpha p = RT$$

we obtain

$$\Delta\alpha p + \alpha\,\Delta p = R\,\Delta T$$

which substituted in Eq. (4) (page 51) gives

$$-c_v\,\Delta T = R\,\Delta T - \alpha\,\Delta p$$

or

$$(R + c_v)\,\Delta T = \alpha\,\Delta p$$

Now, the gas constant R is the difference between the specific heat of air at constant pressure c_p and the specific heat of air at constant volume c_v. Therefore, $R + c_v = c_p$, and the above equation reduces to

$$c_p\,\Delta T = \alpha\,\Delta p$$

This equation shows that the temperature varies in direct proportion to the variation in pressure.

To obtain a more convenient expression for the temperature variation in an adiabatic process, we substitute for Δp from the equation of static equilibrium, and obtain

$$\frac{\Delta T}{\Delta z} = -\frac{g}{c_p}$$

Substituting the numerical values for the constants g and c_p, we obtain

$$\frac{\Delta T}{\Delta z} = -1°\text{C. per 100 m. ascent}$$

In other words, nonsaturated air cools at a rate of 1°C. per 100 m. when it ascends dry-adiabatically. This is called the *dry-adiabatic rate of cooling* or the *dry-adiabatic lapse rate.* Conversely, if the air descends adiabatically, the temperature will increase 1°C. per 100 m. descent. Expressed in degrees Fahrenheit and feet, the dry-adiabatic lapse rate is approximately 1°F. per 180 ft., or 5.6°F. per 1000 ft.

Ascent and Descent of Saturated Air.—We consider next air that is saturated with moisture and that ascends through the atmosphere without heat being supplied to or withdrawn from it. Owing to expansion alone, the air would cool at a rate of 1°C. per 100 m. However, when saturated the air cools, condensation takes place, and the latent heat of evaporation is liberated and tends to heat the air. The two processes counteract one another, but the cooling due to expansion is the major one, with the result that saturated air cools while it ascends, but at a slower rate than does nonsaturated air. The rate of cooling of ascending saturated air is called the *moist-adiabatic rate of cooling.*

It is readily seen that the moist-adiabatic rate of cooling varies with the temperature of the air. To prove this, we let saturated air ascend 100 m. The cooling due to expansion is then 1°C. From Fig. 9, we see that a temperature decrease of 1°C. will result in the condensation of a considerable amount of moisture when the air temperature is high, for the saturation curve is very steep at high temperatures. If the air temperature is low, a cooling of 1°C. results in the condensation of a very small amount of moisture. Consequently, the amount of liberated heat per 100 m. ascent is much greater at high temperatures than at low temperatures; and, therefore, the moist-adiabatic lapse rate is smaller at high temperatures than at low temperatures. At extremely low temperatures, the amount of moisture is so small that the moist-adiabatic lapse rate is almost equal to the dry-adiabatic rate. At high temperatures the moist-adiabatic lapse rate is approximately one-half of the dry-adiabatic rate, *viz.*, about 0.5°C. per 100 m., or about 3°F. per 1000 ft.

The above principles are illustrated in Fig. 34. A unit of air at *A* ascending through the atmosphere would cool at the dry-adiabatic rate of 1°C. per 100 m., or 10°C. per kilometer, until it becomes saturated at, say, *B*. From there on, it cools at the moist-adiabatic rate as indicated by the line *BC*. Another unit of air at A' will cool dry-adiabatically until it becomes saturated at B' and then cool moist-adiabatically as indicated by the line $B'C'$. Note that at low temperatures (line $A'B'C'$) the moist- and dry-adiabats are almost coincident, whereas at higher temperatures (line ABC) the angle between the two adiabats is much greater. This is due to the fact that at low temperatures the air contains so little water vapor that only small amounts of water vapor can be condensed. A diagram of this kind may be used to evaluate the temperature variations of air masses that move up or down through the atmosphere.

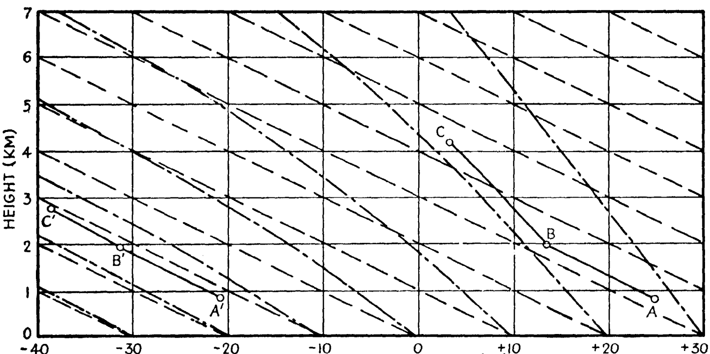


FIG. 34.—Dry-adiabats (broken lines) and moist-adiabats (broken-dotted lines) in a temperature-height diagram.

The Adiabatic Chart.—Instead of using the temperature-height diagram shown in Fig. 34, meteorologists find it more convenient to use a diagram in which temperature (or some function of temperature) is the abscissa and pressure (or some function of pressure) is the ordinate. A diagram of this kind is shown in Fig. 35. It represents a graphical solution of the first law of thermodynamics.

The vertical lines indicate temperature, and the almost horizontal lines indicate atmospheric pressure. The broken lines are dry-adiabats; the broken-dotted lines are moist-adiabats. Note that the moist-adiabats deviate much from the dry-adiabats at high temperatures and approach the dry-adiabats at low temperatures.

In addition to the dry- and moist-adiabats, the diagram contains a set of dotted lines which indicate the specific humidity (see page 19) of saturated air.

To illustrate the use of this diagram, we shall consider some examples.

1. Suppose that we have observed atmospheric pressure 1000 mb., air temperature 26°C., and relative humidity 75 per cent.

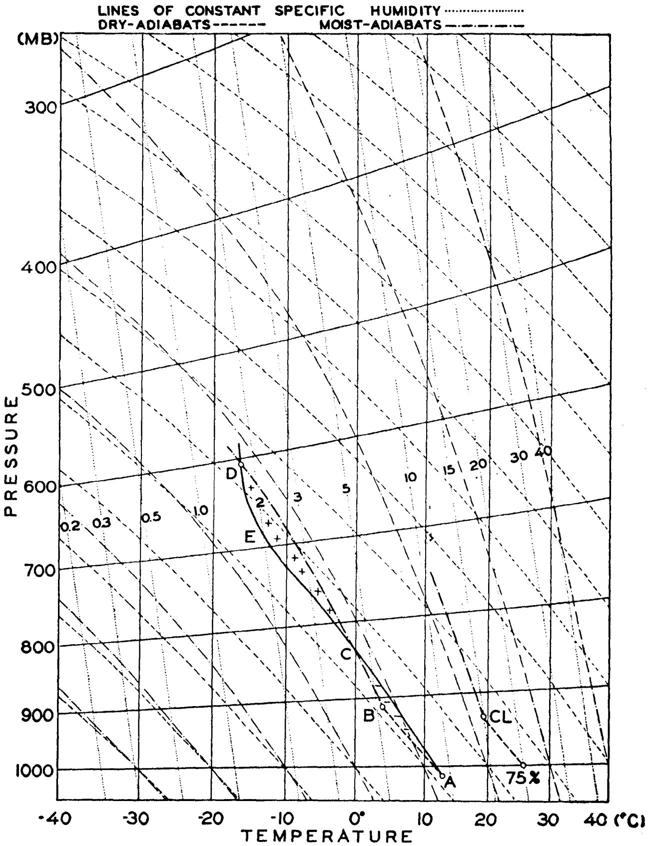


FIG. 35.—Adiabatic chart (the Refsdal Aerogram).

Plot in Fig. 35 the point whose temperature is 26°C. and whose pressure is 1000 mb. If the air were saturated, its specific humidity would be 20 grams, since the 20-gram line goes through this point. However, since the relative humidity is 75 per cent, the specific humidity will be 20 × 0.75 = 15 grams. The dotted line marked 15 then indicates the specific humidity of the air in this case.

If the unit of air moved upward through the atmosphere, it would cool, and the point in the diagram that represents the unit would follow the dry-adiabat through the point represented by 1000 mb. and 26°C. Follow the dry-adiabat through this point to intersection with the 15-gram humidity line. Here, the specific humidity of the rising air would be equal to the saturation value; the air has then reached its *condensation level CL*. If the air ascends farther, it will cool moist-adiabatically, the point representing the unit will move along the moist-adiabat, and, as it does so, water vapor will condense.

Thus, the adiabatic chart is a useful tool by means of which the meteorologist can evaluate the physical changes of air masses that move up or down through the atmosphere.

2. Suppose next that an aircraft carrying a meteorograph has measured the distribution of pressure, temperature, and humidity in the air column. The corresponding values of pressure and temperature are plotted on an adiabatic chart, as shown in Fig. 35 by the curve *ACED*. Suppose that the unit of air at *A* is moved upward; it will then follow the dry-adiabat to its condensation level *B* and from there follow the moist-adiabat *BCD*. Now, the curve *ACED* represents the conditions of the air column, and the curve *ABCD* represents the conditions of the unit of air at *A* while it moves upward. It will then be seen that the rising unit of air is colder than the environment while it moves from *A* to *C*. Since, for any given pressure, colder air is denser than warmer air, it follows that the rising unit would be colder than the environment below the level *C*; it would therefore tend to resist upward motion; and if it were left to itself, it would sink back to the level whence it came. However, if the unit at *A* were forced to ascend beyond the level *C*, it would be warmer than the environment; it would then be less dense than the environment and would therefore tend to rise farther on account of the buoyancy forces. At the level *D*, it would again have the same density as the environment. The adiabatic chart may therefore be used for determining the ease with which vertical motion can be established in the air column.

Since energy is required to create motion, it is of interest to determine the amount of energy available. This question, too, is answered by the use of the adiabatic chart. The area ABC in Fig. 35 expresses the amount of energy required to make a unit mass of air at A ascend to the level C. This area, which is called the *negative area,* expresses the resistance against lifting. The area CDE in Fig. 35 expresses the amount of energy that can be released by a unit mass at A if it were lifted beyond the level C. This area is called the *positive area.* The difference between the positive area and the negative area expresses the amount of *available energy.* In the case considered in Fig. 35 the positive area is larger than the negative area, and energy is available for the development of vertical currents. Whether or not energy is available depends entirely on the lapse rate and the moisture content of the air column. We are thus led to consider the stability conditions of the air; and, in doing this, it is necessary to bear in mind the following definitions:

1. The *lapse rate of temperature* is the rate at which the temperature in the air column decreases along the vertical.
2. The *dry-adiabatic lapse rate* is the rate at which nonsaturated air cools per unit distance (say, 100 m.) when it ascends through the atmosphere.
3. The *moist-adiabatic lapse rate* is the rate at which saturated air cools per unit distance (say, 100 m.) when it ascends through the atmosphere.
4. The *condensation level* is the level at which air that ascends dry-adiabatically becomes saturated.

To facilitate the following discussion, we introduce the symbols

$$\begin{aligned}\gamma &= \text{the lapse rate of the air column}\\ \gamma_d &= \text{the dry-adiabatic lapse rate}\\ \gamma_m &= \text{the moist-adiabatic lapse rate}\end{aligned}$$

Potential Temperature.—It is evident from the discussion in the foregoing paragraph that the air suffers considerable temperature variations when it partakes in adiabatic processes. In many cases, it is of interest to compare the temperatures of air masses at different levels. But since the temperature varies with the pressure, it is necessary to determine the temperatures that the air would have if it were brought adiabatically down to a standard level, or, more correctly, to a standard pressure. We choose 1000 mb. as the standard pressure and define the *potential temperature* as the temperature that the air at any level would have if it were brought dry-adiabatically down to the level where the pressure is 1000 mb.

The potential temperature is obtained from an adiabatic chart. Let the curve PQ in Fig. 36A represent the temperature distribution in the air column. To obtain the potential temperature of the point P, follow the dry-adiabat through P to intersection with the line indicating 1000 mb. The temperature line through this point indicates the potential temperature of the point P in Fig. 36A. In exactly the same manner, we determine the potential temperature of the point Q and find that its potential temperature is higher than that of the point P. In other words, the potential temperature in Fig. 36A increases with elevation.

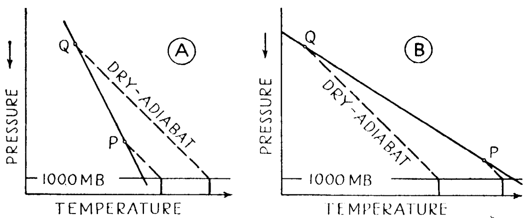


FIG. 36.—Showing how to obtain the potential temperature.

We determine next the potential temperature of the points P and Q in Fig. 36B and find that the potential temperature decreases along the vertical. The essential difference between the two cases is that in Fig. 36A the lapse rate of temperature in the air column is less than the dry-adiabatic rate and in Fig. 36B it is larger than the dry-adiabatic rate. We thus see that the following rules hold:

1. The potential temperature in the air column increases with elevation when the lapse rate is less than the dry-adiabatic rate, or when $\gamma < \gamma_d$.
2. The potential temperature in the air column is constant with elevation when the lapse rate is equal to the dry-adiabatic rate, or when $\gamma = \gamma_d$.
3. The potential temperature in the air column decreases with elevation when the lapse rate is greater than the dry-adiabatic rate, or when $\gamma > \gamma_d$.

When a unit of nonsaturated air moves up or down through the atmosphere, it follows a certain dry-adiabatic line in the adiabatic diagram. The point of intersection between this dry-adiabatic line and the 1000-mb. pressure line is stationary during this process. It follows then that *the potential temperature of a unit of air is constant when the air moves dry-adiabatically up or down through the atmosphere.* Thus the potential temperature is *conservative* with respect to dry-adiabatic processes, and ascending and descending air masses may be identified by their potential temperature.

When condensation occurs, the air moves moist-adiabatically, and the potential temperature increases on account of the liberation of the latent heat of vaporization. Furthermore, near the earth's surface, the air receives or gives off heat to the underlying surface; the process is then no longer adiabatic, and the potential temperature is not conservative. The use of potential temperature as a means of identifying air masses is therefore restricted to the free atmosphere when the air is so far removed from heat and cold sources that the nonadiabatic influences are negligible.

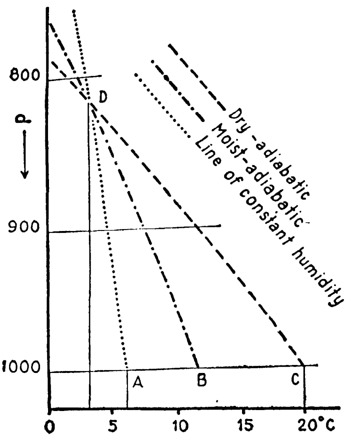


FIG. 37.—The dry-adiabatic variation of dew-point temperature is about one-fifth of the dry-adiabatic variation of temperature. Above the condensation level (D) both temperature and dew point vary moist-adiabatically.

Since no heat is added to or withdrawn from the air that partakes in a dry-adiabatic process, the entropy is, by definition, constant. Thus a surface of constant potential temperature in the free atmosphere is also a surface of constant entropy, or an *isentropic surface.* When the air moves dry-adiabatically, it must be found at successive intervals in the same isentropic surface. Mostly through the work of Rossby in recent years, maps showing the distribution of the meteorological elements in isentropic surfaces are now in general use as an aid in identifying air masses from one day to the next. This method of analysis, which is called *isentropic analysis,* has led to considerable improvement in our methods of forecasting weather phenomena.

Evaluation of the Condensation Level.—The condensation level of air that ascends adiabatically may be determined by means of an adiabatic chart. Let the point C in Fig. 37 represent a unit of air with pressure 1000 mb. and temperature 20°C. This unit has a certain specific humidity represented by the dotted line AD. Follow the dry-adiabat through C to intersection with the line of constant humidity. If the air ascended dry-adiabatically, it would be saturated at the level D, which therefore represents the condensation level of the unit of air at C.

The following relation between specific humidity and dew point is of interest: If the unit of air at C were cooled at constant pressure and with constant moisture content, it would become saturated at the point A of which the temperature, by definition, is the dew point of the unit at C. On the other hand, if the unit at C ascended dry-adiabatically with constant moisture content, its dew point would vary along the line AD.

If, now, the dew-point temperature is observed directly, the condensation level of the unit at C may be determined in the following manner: Follow the dry-adiabat through C and the humidity line through A (the dew point) to intersection at D, which determines the condensation level of the unit at C.

It will be seen from Fig. 37 that the unit at C, which ascends to D, cooled dry-adiabatically about 17°C. Simultaneously, the dew point of the ascending unit decreased only 3.5°C. Therefore, the dry-adiabatic variation of the dew point is only one-fifth, of the dry-adiabatic variation of the temperature. Since the dry-adiabatic lapse rate is 1°C. per 100 m., it follows that the dry-adiabatic variation in dew point is 0.2°C. per 100 m. ascent. Therefore, when a unit of air ascends, *the air temperature overtakes the dew point at a rate of about* 0.8°*C. per* 100 *m. ascent, or about* 1°*F. per* 230 *ft. ascent.*

Suppose now that the air temperature is 75°F. and the dew point is 65°F. Applying the above rule, we find that the condensation level is then 2300 ft. above the station. The above relations are not exact, but they are sufficiently accurate for all practical purposes. The fact that the computed condensation level is 2300 ft. means not necessarily that clouds will form at that level but that if the air ascended dry-adiabatically the base of the clouds would be found at that level. Whether or not the air will ascend depends entirely on the stability conditions of the air columns.

CHAPTER V

STABILITY AND INSTABILITY

Most weather phenomena depend on whether the air masses are stable or unstable. When the air is stable, vertical motion is suppressed, whereas, when it is unstable, vertical currents develop; the air masses at high levels are then "too heavy" to be supported by the potentially warmer air masses at low levels. This results in vertical air currents. The process of overturning of unstable air masses is called *convection.* The weather phenomena typical of convection are cumulus or cumulo-nimbus clouds, showers, squalls, thunderstorms, hail, wind gusts, bumpiness, etc.

Definitions.—To test whether the air column is stable or unstable, we let a unit of air be displaced a small distance up or down. If the unit, after being thus displaced, has a tendency to return to its original level, the equilibrium is said to be *stable.* However, if the unit after being displaced has a tendency to move farther away from its original level, the equilibrium is said to be *unstable.* The limiting case may occur, *viz.*, when the displaced unit tends neither to return to nor to move farther away from its original level; in this case, the equilibrium is said to be *neutral* or *indifferent.*

A rocking chair at rest is in a stable state of equilibrium, for if it were rocked slightly it would return toward its original position. A pencil balanced on its point is in an unstable state of equilibrium, for the slightest impulse would cause it to move farther away from its original position. A perfectly round ball on a highly polished and perfectly horizontal slate is in an indifferent state of equilibrium.

Since any minute disturbance will upset the unstable systems and bring them to a stable state, it follows that unstable systems cannot exist for any appreciable length of time. The transition from unstable to stable states of equilibrium involves a reduction of the potential energy, and all systems left to themselves will tend to avoid instability and obtain a minimum of potential energy.

The principles of stability and instability that hold for solid bodies also apply to the atmosphere, but the conditions are here more complicated, owing to the fact that the air is compressible and therefore changes its density as it moves upward or downward. A second complication arises when the air becomes saturated with moisture. The latent heat of varporization is then liberated and is used to heat the air; this heating affects the density and therefore the stability of the air, also.

In order to derive a convenient criterion for the stability and instability of air masses, we shall resort to the adiabatic chart and discuss nonsaturated and saturated air separately.

Nonsaturated Air.—Figure 38 shows a simplified adiabatic chart on which are plotted two hypothetical temperature curves. The broken line indicates the dry-adiabat. We consider first curve 1. Suppose that the unit of air at P is displaced upward; the point in the diagram that represents the unit will then follow the dry-adiabatic through P. It is readily seen that the displaced unit would be colder and therefore denser than the environment. It would therefore sink down toward the level whence it came. By definition, then, the stratification is a *stable* one.

We next consider curve 3 where the temperature decreases more rapidly with elevation. If the unit at P is displaced upward, it would cool dry-adiabatically as before, but it would now be warmer than the environment and would therefore rise farther on account of its being less dense than the surrounding air. In this case, the air is *unstable*.

As a limiting case, it might happen that the temperature curve coincided with the dry-adiabat (curve 2, Fig. 38). In this case, the unit at P, after being displaced upward, would have exactly the same temperature as the environment; it would therefore have the same density as the environment, and it would be at rest after the displacement. In this case the air is in a *neutral*, or *indifferent*, state of equilibrium.

In the above examples, we have tested the stability conditions by displacing an arbitrary unit of air a small distance upward and have then compared the density of the unit with the density of the surrounding air. The same results are obtained if we displace the unit downward. If the temperature-height curve is as shown by curve 1, the unit would return to its original level; on the other hand, if the temperature distribution were as indicated by curve 3, the unit would move farther away from its original level; furthermore, if the temperature distribution were as indicated by curve 2, the unit would not move farther away from its original level, if it were displaced a small distance upward or downward.

The essential difference between curve 1 and curve 3 in Fig. 38 is that in the former the lapse rate of temperature in the air column is less than the dry-adiabatic rate, whereas, in the case of

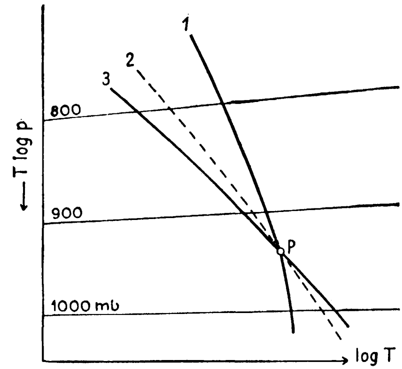


FIG. 38.—Stability conditions of dry air; broken line indicates the dry-adiabat. Note that the greater the slope of the temperature curve, the smaller the lapse rate.

curve 3, it is larger than the dry-adiabat, and this distinguishes the stable from the unstable condition. We may therefore write, for nonsaturated air,

1. The air column is stable when its lapse rate is smaller than the dry-adiabatic rate, or when $\gamma < \gamma_d$.

2. The air column is indifferent, or neutral, when its lapse rate is equal to the dry-adiabatic rate, or when $\gamma = \gamma_d$.

3. The air column is unstable when its lapse rate is larger than the dry-adiabatic rate, or when $\gamma > \gamma_d$.

These criteria hold regardless of the relative humidity as long as the air is not saturated with moisture.

Saturated Air.—The conditions for saturated air are similar to those for nonsaturated air, except that the moist-adiabat is substituted for the dry-adiabat. To show this, we consider Fig. 39 where the broken-dotted line represents the moist-adiabat.

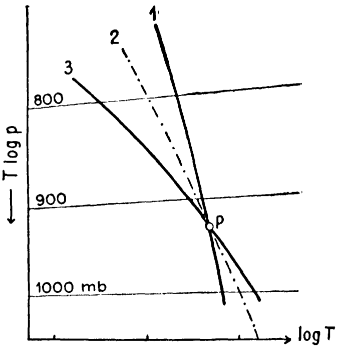


FIG. 39.—Stability conditions of saturated air; broken-dotted line indicates the moist-adiabat.

The air that is displaced upward will cool at the moist-adiabatic rate. If the temperature distribution in the air column is as shown by curve 1, the ascending air will be colder than the environment; if the temperature distribution is as shown by curve 3, the rising air will be warmer than the environment. Proceeding exactly in the same way as that described in the previous section, we obtain the following criteria for saturated air:

1. The air column is stable when its lapse rate is less than the moist-adiabatic rate, or when $\gamma < \gamma_m$.

2. The air column is indifferent, or neutral, when its lapse rate is equal to the moist-adiabatic rate, or when $\gamma = \gamma_m$.

3. The air column is unstable when its lapse rate is larger than the moist-adiabatic rate, or when $\gamma > \gamma_m$.

Conditional Instability.—In the two foregoing sections, we discussed the stability conditions of an air column when the ascending air was either saturated or nonsaturated. In nature, it happens often that the ascending air passes its condensation level, with the result that the air becomes saturated while it ascends. In such cases, the air cools dry-adiabatically below the condensation level and moist-adiabatically above it.

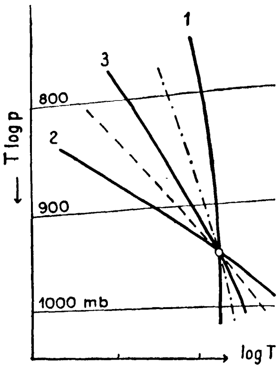


FIG. 40.—Absolute stability (1), absolute instability (2), and conditional instability (3) in relation to moist- and dry-adiabatic lines.

To obtain stability criteria in this case, we consider Fig. 40, in which are plotted three hypothetical temperature curves 1, 2, and 3. Curve 1 is characterized by a lapse rate that is less than the moist-adiabatic rate (*i.e.*, $\gamma < \gamma_m$). It is readily seen from Fig. 40 that if a unit of air is made to ascend it would be colder than the surrounding air whether or not condensation occurs. In this case, we say that the air column is *absolutely stable*.

We next consider curve 2, which is characterized by a lapse rate greater than the dry-adiabatic (*i.e.*, $\gamma > \gamma_d$). Displacing the unit upward, we see that it will be warmer than the surrounding air whether or not saturation occurs. The displaced unit will therefore be accelerated upward, and the air column is *absolutely unstable*.

The most frequently occurring case is a lapse rate in the air column that is larger than the moist-adiabatic and smaller than the dry-adiabatic rate, or $\gamma_d > \gamma > \gamma_m$. This case is represented by curve 3 in Fig. 40. In this event, the air column may be stable or unstable, depending on the distribution of moisture.

To discuss this case in greater detail, we return to Fig. 35 and consider the unit of air at A. If it were lifted, it would cool dry-adiabatically to its condensation level B above which it would cool moist-adiabatically. Below the level C, it would be colder than the surrounding air, and it would resist lifting. However, if it were made to ascend above the level C, it would be warmer than the environment and would therefore ascend farther. In other words, the air column, in this case, is stable with respect to small impulses; but if the impulse is sufficiently strong to bring the unit beyond the level C, instability would result.

Now, if the unit at A had a high relative humidity, its condensation level would be low. The negative area ABC would then be small, and the positive area CDA would be large. On the other hand, if the relative humidity were low, the condensation level would be high; the negative area would increase and the positive area decrease. Furthermore, if the air were exceedingly dry, the positive area would vanish, and there would be nothing but a negative area. As explained on page 57, only when the positive area is larger than the negative area is there energy available for the creation of vertical currents.

The case where $\gamma_d > \gamma > \gamma_m$ is called *conditional instability*, because the stability is conditioned by the humidity distribution. The state of conditional instability may be subdivided into three classes according to the presence or absence of a positive area and according to whether the positive area is larger or smaller than the negative area.

The foregoing discussion may be summarized as follows:

1. The air column is *absolutely stable*, whether or not saturation occurs, when its lapse rate is less than the moist-adiabatic rate, or when $\gamma < \gamma_m$.

2. The air column is *absolutely unstable*, whether or not saturation occurs, when its lapse rate is larger than the dry-adiabatic rate, or when $\gamma > \gamma_d$.

3. The air column is *conditionally unstable* when its lapse rate is larger than the moist-adiabatic rate and smaller than the dry-adiabatic rate, or when $\gamma_d > \gamma > \gamma_m$. This case may be subdivided into three classes,

a. The real latent type, characterized by a positive area larger than the negative area.

b. The pseudo-latent type, characterized by a positive area smaller than the negative area.

c. The stable type, characterized by the absence of a positive area.

It is of interest to note that a nonsaturated air column that is conditionally unstable is not actually unstable but that instability may be released if the air were lifted to the level C in Fig. 35.

Convective Instability.—It happens quite frequently that the air column is stable but that if the whole air mass is lifted bodily so much that condensation occurs, it becomes unstable. Whether the lifted air will retain its stability or develop instability depends entirely on the distribution of moisture along the vertical.

To prove this, we consider two cases as shown in Fig. 41A and B. In both cases the curve ab represents the distribution of temperature within the layer of air ab. It will be seen that the lapse rate within the layer is less than the moist-adiabatic rate. The layer of air is therefore absolutely stable. If the whole layer is lifted, it will either retain its stability or change into instability, depending on the distribution of moisture. In Fig. 41A the air at b is much dryer than the air at a. If the layer is lifted as a whole, the point a would soon reach its condensation level; and if it were lifted farther, it would cool at the moist-adiabatic rate. In the meantime the unit at b, being drier, will cool at the dry-adiabatic rate until its condensation level b' is reached. By the time the whole layer has been lifted so much that it has become saturated, the temperature distribution within the layer will be as shown by the curve $a'b'$. It will then be seen that the lapse rate is now greater than the moist-adiabatic rate; and since the air is saturated, it is unstable.

We next consider Fig. 41*B*. Here the upper portion of the layer is moister than the lower portion. Consequently, the point *b* reaches its condensation level before the point *a* becomes saturated. When the whole layer has become saturated, the temperature distribution would be as shown by the curve *a′b′*. It will now be seen that the air has become more stable than it was originally.

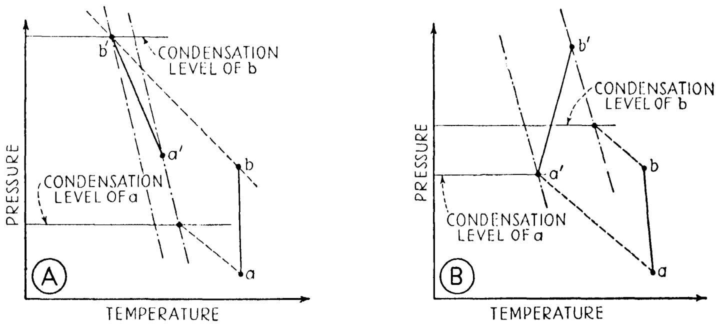


FIG. 41.—Illustrating the meaning of convective instability (*A*) and convective stability (*B*).

The principal difference between the two cases is that, in case *A*, the condensation level of the uppermost unit falls on a moist-adiabat that is *to the left* of the moist-adiabat through the condensation level of the lowermost unit, whereas, in case *B*, the reverse is true. Case *A*, then, represents convective instability, and case *B* represents convective stability.

Since the wet-bulb temperature in an adiabatic process varies along the moist-adiabat, we may say that *the air column is convectively unstable when the decrease in wet-bulb temperature with elevation exceeds the moist-adiabatic rate.*

The importance of convective instability lies in the fact that stable air masses may become unstable through vertical motion of the air mass as a whole. Such motion takes place on the windward side of mountain ranges and also along frontal surfaces. We shall return to the discussion of the weather phenomena caused by instability in later chapters.

CHAPTER VI

TEMPERATURE VARIATIONS AND THEIR RELATION TO THE WEATHER PHENOMENA

In the two preceding chapters, we have discussed the adiabatic temperature changes. In these processes, no heat is supplied to or withdrawn from the air, the temperature variations being entirely due to expansion and contraction as the air comes under lower or higher pressure.

Temperature changes may also be due to actual heat being supplied to or removed from the air. Such temperature changes are called *nonadiabatic.*

Sources of Heat.—Observations show that the temperature increases as we go down through the solid earth, at a rate of about 1°F. per 70 ft. The earth's crust is, however, a poor conductor of heat. As a result, the amount of heat that is conducted through the earth's crust and made available to the atmosphere is negligible. So is also the amount of heat that the atmosphere receives from the stars. The sun may thus be regarded as the only source of the heat energy that is supplied to the earth's surface and the atmosphere. The ultimate cause of all changes and motions in the atmosphere may, therefore, be sought in the energy radiated from the sun.

Radiation.—All bodies, whatever their temperatures, throw off radiation in the form of electromagnetic waves which travel with a speed of about 186,000 miles per second. What we call visible light is only a comparatively narrow band of the entire spectrum of radiation. The hotter the body, the more intense is the radiation, and the character of the radiation varies with the temperature. Thus, if a cold piece of metal is heated, it first throws off radiation which is invisible to the human eye but may be felt as heat. This invisible radiation at relatively low temperatures is characterized by long wave length; it is often referred to as "low-temperature radiation" or infrared radiation, because the wave length is longer than the wave length of visible red. As the temperature increases, the metal begins to glow with a dull red color (visible radiation), and with further heating the color brightens until ultimately the metal becomes white. These changes in color are due to the wave length of the radiation becoming gradually shorter. Radiation in the short wave length is therefore often referred to as "high-temperature radiation."

Now, the temperature of the sun's surface is about 6000°C., whereas the average temperature of the earth's surface is about 20°C. The earth's radiation is therefore invisible (low-temperature, or long-wave, radiation), whereas the sun radiates mainly in the visible range (high-temperature, or short-wave, radiation).

The direct radiation of the sun is absorbed only to a slight extent in the atmosphere because the air is almost transparent to high-temperature radiation. About 43 per cent of the incoming radiation is reflected back to space, about 40 per cent is absorbed by the earth's surface, and the remaining 17 per cent is absorbed by the atmosphere. This means that only 57 per cent of the incoming solar radiation is thermally effective.

The thermally effective 57 per cent of the incoming radiation must be remitted by the earth and the atmosphere as low-temperature radiation. Part of this radiation is absorbed by the clouds, and a considerable portion of it is absorbed by the water vapor in the air. Thus, the water vapor in the air acts as the glass in a greenhouse: it lets through practically all incoming short-wave radiation from the sun, and it tends to prevent the outgoing long-wave radiation from the earth from getting back to the universe.

Transfer of Heat.—Since the atmosphere is a poor absorber and the earth's surface is a good absorber of incoming radiation, the atmosphere receives most of the heat energy via the earth's surface. The heat received in one place may be transported to other places by conduction, radiation, turbulence, and advection.

The transfer of heat through molecular conduction is exceedingly small and may be disregarded in the discussion of atmospheric conditions. Although the radiation transfer of heat is about ten thousand times greater than the molecular transfer, it is still small as compared with the transfer of heat caused by the motion of the air.

The wind is never a steady current; it consists of a succession of gusts and lulls of short period. This irregular motion, which is called *turbulence*, is made up of a number of small eddies that travel with the general air current, superimposed on it. These eddies carry heat, moisture, dust, etc., with them, as they travel from one place to another. The turbulent, or eddy, transfer of heat is most effective in the vicinity of the earth's surface where eddies readily form on account of roughness and the unequal heating and cooling of the ground. In this way the heat absorbed by the earth's surface is distributed through the air column.

Heat may also be transported from one place to another by *advection*, or large-scale air currents. Since these are mainly horizontal currents, the advective transfer of heat is mainly in the horizontal direction, whereas turbulence and convective currents transport heat mainly along the vertical. Though the turbulent and the radiative transfer of heat tend to smooth out temperature contrasts, the advective transfer may create or destroy temperature contrasts, depending on whether the large-scale air currents are convergent or divergent.

To sum up, we may say that the transfer of heat due to molecular conduction is negligible. The radiative transfer of heat is small and may, at most, amount to 4°F. in 24 hr. The main transfer of heat in the atmosphere is brought about by turbulence and convective currents (along the vertical) and the large-scale air currents (in the horizontal direction).

Vertical Mixing.—It was explained in the previous section that turbulence has a tendency to redistribute the heat, moisture, etc., through mixing of neighboring air masses. This mixing takes place partly in the horizontal and partly in the vertical direction. We may therefore speak of vertical and horizontal mixing. The difference between these mixing processes is that horizontal mixing takes place at constant pressure, whereas the air that is involved in vertical mixing is subject to pressure changes as it moves up or down through the atmosphere.

To illustrate the effect of vertical mixing on the distribution of temperature and humidity along the vertical, we consider first a vessel partly filled with cold and salt water, on top of which is a layer of warm and fresh water. The cold and salt water is heavier than the warm and fresh water, and the stratification is a stable one. If turbulent motion is created and maintained in the vessel, the two layers will mix into a homogeneous liquid showing constant temperature and constant salinity along the vertical.

In the atmosphere, the conditions are somewhat different owing to the fact that the air is compressible. The ascending air will be cooled adiabatically, and the descending air will be heated adiabatically. The result of turbulent mixing along the vertical is to create dry-adiabatic lapse rates if the air is nonsaturated and a moist-adiabatic lapse rate if the air is saturated.

The effect of vertical mixing on the temperature distribution is illustrated in Fig. 42*A* and *B*. It will be seen that in case *A* the lapse rate in the air column is less than the dry-adiabat; the stratification is stable. Through mixing the lapse rate will

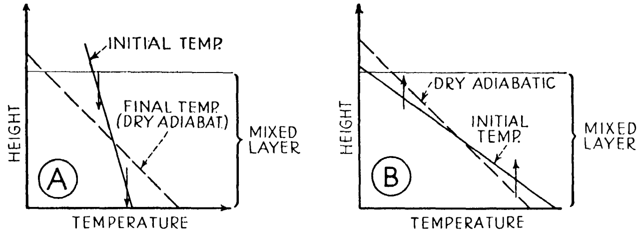


FIG. 42.—Showing the influence of vertical mixing on the temperature distribution under stable (*A*) and unstable conditions (*B*).

approach the dry-adiabat; and if no heat is added to or withdrawn from the air, the mean temperature at the end of the process must be the same as at the beginning of the process. This implies that heat must be transported from the top to the bottom of the mixed layer.

In case *B* the lapse rate is greater than the dry-adiabatic rate. Since the final result of vertical mixing is to create an adiabatic lapse rate, it follows that heat must be transported from the bottom to the top of the mixed layer.

We shall next consider the influence of vertical mixing on the humidity distribution. The specific humidity of the air, being independent of the adiabatic changes, will mix in the same way as the salinity in the example described at the beginning of this section. Therefore, after complete stirring, the specific humidity is constant along the vertical.

Normally, the atmosphere is stably stratified (Fig. 42*A*), and the specific humidity decreases with elevation. Now, if the wind increases, the turbulent mixing will increase, and the temperature will change so as to conform with the dry-adiabat. Simultaneously, the specific humidity will change so as to become constant along the vertical. Thus, under normal conditions, vertical mixing will tend to decrease the temperature and increase the moisture content in the upper portion of the mixed layer and increase the temperature and decrease the moisture content in the lower portion of the mixed layer. This, again, will tend to decrease the relative humidity near the earth's surface and increase the relative humidity in the upper portion of the mixed layer.

A typical example is shown in Fig. 43. Initially, the temperature is constant along the vertical and the air is saturated

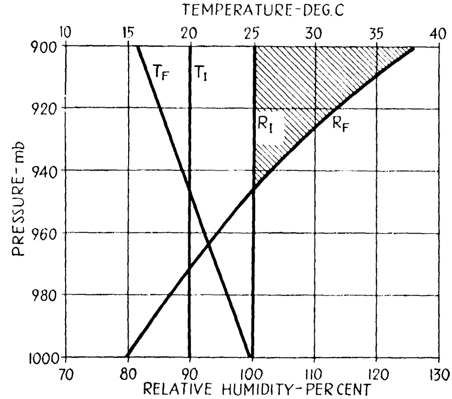


FIG. 43.—T_I and R_I represent the initial distribution of temperature and relative humidity. T_F and R_F represent the final distribution after complete vertical mixing.

throughout. After complete stirring, the relative humidity has decreased in the lower portion and increased in the upper portion of the mixed layer. In the absence of condensation nuclei, the air in the upper portion of the layer would be supersaturated (relative humidity exceeding 100 per cent). However, since condensation nuclei are always present in abundant amounts, the superfluous water will condense above the level where the relative humidity is 100 per cent. This level is called the *mixing condensation level.*

The turbulence set up by the roughness of the ground has a marked tendency to prevent fogs from forming and to dissolve fogs that form under adverse conditions. On the other hand, the turbulent mixing has a marked tendency to cause condensation to occur in the upper portion of the layer influenced by the roughness of the ground. This layer, which is called the friction layer, is about 2000 to 4000 ft. deep. In the upper portion of this layer, condensation phenomena are relatively frequent, whereas they are relatively rare near the ground. From the point of view of aviation, this is highly favorable, for turbulence has a marked tendency to create a cloudless layer next to the earth's surface.

Since the turbulence and vertical mixing increase with the wind velocity, it follows that fog will not form in strong wind. On the other hand, when the wind is strong, clouds tend to form at some distance above the ground. The clouds that form under these conditions are of the stratus type.

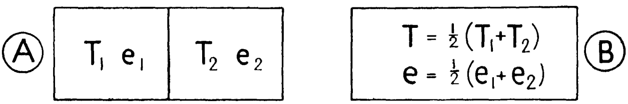


FIG. 44.

Horizontal Mixing.—The mixing that occurs between horizontally adjacent air masses may be assumed to take place at constant pressure; there is then no adiabatic change involved. To simplify, we consider two equal masses of nonsaturated air with temperatures T_1 and T_2 and vapor pressures e_1 and e_2, respectively. Suppose that the wall in Fig. 44*A* is removed and the masses mixed completely. The temperature and the vapor pressure of the mixture would then be as shown in Fig. 44*B*.

Now, the saturation vapor pressure varies with the temperature as shown in Fig. 45.[1] Plotting the vapor pressure versus temperature in this diagram, we obtain the point *A* for the colder component and the point *B* for the warmer component. The physical conditions of the mixture would then be represented by the point *C*, which is halfway between the points *A* and *B*. It will then be seen that two different air masses of widely different temperatures, neither of which is saturated, might become saturated or supersaturated after complete horizontal mixing.

In the absence of condensation nuclei, the point *C* would fall to the left of the saturation curve, supersaturation being thus indicated. However, since condensation nuclei are always present in abundant amounts, the superfluous water will condense.

[1] See also Fig. 9.

When condensation occurs, the latent heat of vaporization is liberated and the air is heated to a certain extent; this tends to reduce the amount of condensed water. Without condensation, the point C would have the temperature T, and it would be supersaturated by the amount CD. With condensation the temperature would rise from T to T_s, and the air would be saturated. The conditions of the mixture would be represented by the point E in Fig. 45, and the superfluous water, represented by the line CG, would condense.

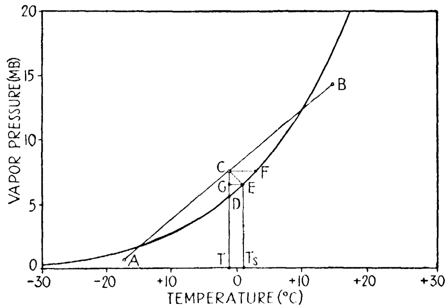


FIG. 45.

This example shows that it is possible to produce condensation of water vapor by mixing (horizontally) two air masses of different temperature. The actual amount of condensed water can be computed, and it will then be found that the amount is exceedingly small for such temperature contrasts as occur in the atmosphere. Moreover, since horizontal and vertical mixing operate together and since the effect of the latter is so much greater than that of the former, it is evident that mixing cannot produce condensation near the earth's surface. However, condensation will often occur in the upper portion of the mixed layer. We shall return to this discussion in later sections on stratus and fog.

Heating and Cooling of Air over Land.—The amount of radiation absorbed by the earth depends on its color or its albedo.[1] The portion that is absorbed is partly used to evaporate water and partly expended in chemical processes, but the major part

[1] The reflective power of the earth (*i.e.*, the albedo) varies within wide limits. Thus, clouds and snow-covered ground may reflect as much as 80 per cent of the incoming radiation, whereas black soil, water surfaces, etc., reflect only a relatively small portion of the incoming radiation.

goes to heat the earth. The heat thus absorbed is accumulated in the upper few centimeters of the earth, owing to the very slow conduction of heat in the earth. The air that comes into contact with the earth becomes heated, and the lapse rate in the very lowest layer of air increases rapidly. When the lapse rate surpasses the dry-adiabatic, the layer becomes unstable and vertical currents are set up that carry heat and moisture, picked up from the surface, to higher levels. As the sun gets higher in the sky, the heating increases and the unstable layer of air increases in thickness, the heat obtained from the surface being transported to higher and higher levels. The diurnal heating of the earth's surface results in a steep lapse rate and in stirring of the lower atmosphere.

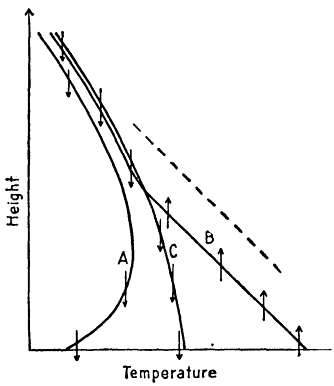


FIG. 46.—Illustrating the diurnal variation in stability over land. *A*. early morning; *B*. midday; *C*. evening; broken line, dry-adiabat. The arrows indicate the direction of the eddy transfer of heat.

As the temperature of the earth's surface increases, the outgoing radiation increases, too; and some two hours after the sun has reached its maximum altitude, there is balance between the loss and gain of heat. The temperature of the earth reaches its maximum and begins to decrease. The cooling of the earth's surface affects the air temperature; and, as before, the influence is greatest near the surface, with the result that the air cools more quickly along the earth than above. The temperature lapse rate decreases, and the air becomes stable again. Figure 46 shows three typical temperature-height curves illustrating the diurnal variation of temperature in the lower atmosphere.

It is worthy of note that the midday temperature on sunny summer days is higher over macadam and sandy fields than over grassland, is lower over wet grounds and woods, and is still lower over water. These differences in temperature give impulses of considerable intensity to convective currents. Local currents of this kind are the main cause of the "bumpiness" experienced by pilots above level ground on warm days.

Heating and Cooling of Air over Oceans.—Although the incoming radiation is absorbed to a considerable extent by the oceans, the temperature of the ocean surface remains almost constant day and night. This condition is due to a variety of causes. Part of the absorbed heat is expended in evaporating water, and the remaining heat is distributed over a deep layer of water, the specific heat of which is very large. Although most of the radiation is absorbed in the surface layer, the mixing caused by wind and waves distributes the heat in a deep layer, with the result that no appreciable variation occurs in the sea surface. The air that is in contact with the sea adapts its temperature to that of the sea surface. Observations show that the diurnal range of the air temperature close to the sea surface is less than 1°F. (0.5°C.).

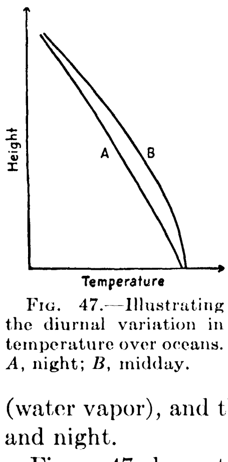


FIG. 47.—Illustrating the diurnal variation in temperature over oceans. *A*, night; *B*, midday.

The regulating influence of the sea surface on the air temperature decreases with elevation. A few hundred meters above the sea, the temperature variation is controlled mainly by radiation, and the air has a diurnal variation in temperature that is greater than the variation at the surface of the sea. At still greater altitudes, the effect of radiation decreases on account of lack of absorbing substance (water vapor), and the temperature remains almost constant day and night.

Figure 47 shows two characteristic temperature-height curves over sea. It is readily seen that the lower part of the atmosphere has a tendency to be stable by day and unstable by night. At greater heights, the reverse is true.

The diurnal variation of stability and instability in the lower atmosphere over oceans is in sharp contrast to the conditions prevailing over land, and this has marked influence on the weather. Over land, all instability phenomena have a maximum frequency and intensity in the afternoon, whereas over oceans the diurnal variation is slight and the maximum frequently occurs at night.

Heating and Cooling of Traveling Air Masses.—An air mass that is colder than the surface over which it travels is called a *cold air mass.* It will be heated from below, and, by continued traveling over a warmer surface, instability will develop in the lower layers and gradually spread upward. The vertical currents resulting from instability will carry heat and moisture to higher and higher levels. The changes in such an air mass are completely analogous to those in air over land that is heated by sunshine on the earth's surface, which we have described in a previous section.

If such an air mass travels over land, it will have the effect of the diurnal temperature changes superimposed on the effect of its travel toward warmer regions. The instability of the air will then vary during the day, having a maximum in the early afternoon and a minimum in the early morning. If the difference in temperature between the surface and the air is large, the air will remain unstable day and night. If the difference is small, the air may become stable in the night.

An air mass that is warmer than the surface over which it travels is called a *warm air mass.* Through continued traveling toward colder regions, it will be cooled from below and will acquire pronounced stability in the lower layers. The stability hinders vertical currents, and the cooling will be limited to the lower layer. If such an air mass travels over land, it will have the diurnal variation in stability superimposed on the effect of its travel toward colder regions. If the temperature difference between the surface and the air is large, it will remain stable day and night. If the difference is small, it will be stable at night and unstable by day.

Over oceans the conditions are different inasmuch as the diurnal variation of stability is small. Over oceans, therefore, stability and instability are determined mainly by the travel of the air masses, and there is but slight diurnal variation. This is particularly true in high and intermediate latitudes.

Figure 48 shows some typical temperature-height curves in traveling air masses under various conditions. When the air is heated from below, the condensation level will tend to rise because the air becomes relatively drier. However, when the heating takes place over oceans, the absorption of moisture is usually sufficient to keep the condensation level relatively low. On the other hand, heating over land results in a relatively high condensation level (see Fig. 48, curves *B* and *C*). When the air is cooled from below, the condensation level lowers. If the wind velocity is slight, the cooling would be greatest near the surface (curve *D*) and fog might form. With high wind, the turbulent mixing would tend to create a normal lapse rate near the earth's surface; the temperature-height curve would be of type *E*, and a layer of low stratus clouds would be the normal occurrence.

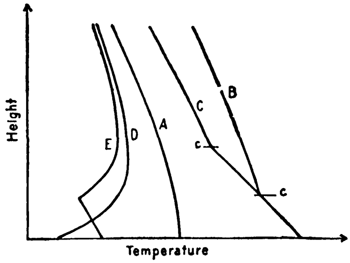


FIG. 48.—Types of temperature-height curves in traveling air masses. *A*, initial curve; *B*, after travel over a warmer ocean; *C*, after travel over a warmer continent; *c*, condensation level; *D*, after travel over a colder surface, slight wind velocity (slight turbulence); *E*, after travel over a colder surface, high wind velocity (strong turbulence).

Convection.—We have seen in the foregoing section that instability is created in the lower layer of the atmosphere either through the diurnal heating of the earth's surface by the sun or through heating of the air when it travels toward warmer regions. A number of phenomena such as gustiness, cumulus clouds, showers, squalls, and thunderstorms are directly caused by instability. We shall therefore discuss the consequences of instability in greater detail.

As soon as the temperature lapse rate near the earth exceeds the dry-adiabatic, the slightest disturbance will upset the stratification. Air from the earth's surface rises, and air from higher levels sinks to replace the ascending masses. This process of overturning of unstable air is called convection. If the rising currents reach the condensation level, clouds will form. The descending air surrounding the rising masses will be heated adiabatically; the relative humidity will be lowered. Thus, the sky will be characterized by broken clouds of the cumulus type, varying from fair-weather cumulus (Fig. 24) and towering cumulus (Fig. 25) to cumulo-nimbus (Figs. 26 and 27).

The ascending currents reach only to the top of the unstable or conditionally unstable layer. The weather phenomena that convection will produce depend on the depth of the unstable layer, the height of the condensation level, and the distribution of temperature aloft. We shall discuss a few typical cases separately.

Case 1.—The condensation level is essentially above the top of the unstable layer. In this case the ascending currents do not reach the condensation level, and no clouds result (dry convection). This case is typical of the conditions in the early morning after clear nights. By continued heating from below, the depth of the unstable layer will grow; and when it builds up to the condensation level, clouds will form. The type of clouds that form will depend on the stability conditions above the condensation level.

Case 2.—We consider next Fig. 49*A*. Near the earth's surface the lapse rate is greater than the dry-adiabatic rate. The stratification is unstable, and the rising air will cool dry-adia-

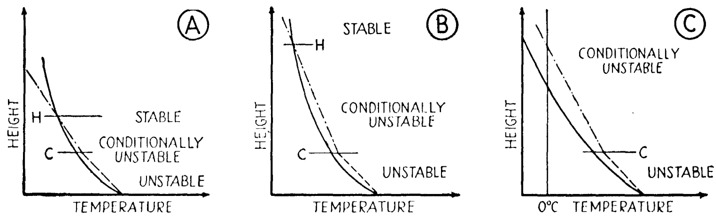


FIG. 49.—Types of stability conditions leading to the formation of *A*, fair-weather cumulus; *B*, towering cumulus; *C*, cumulo-nimbus.

batically until it reaches its condensation level *C* and then moist-adiabatically. It will be seen from the diagram that the ascending air, when it passes the level *H*, will be colder than the surrounding air. The ascending currents, when they pass this level, will be retarded, and they cannot penetrate to greater heights. The base of the clouds will be at the level *C*, and the top will be slightly above *H*. Owing to the pronounced stability above the level *H*, the cumulus clouds will be flat, and they will show no tendency to produce towers or protuberances that grow upward. The type of cumulus that forms under these circumstances is the fair-weather cumulus (cumulus humilis, Fig. 24).

Case 3.—We consider next Fig. 49*B*. The ascending air cools dry-adiabatically to its condensation level *C* and then moist-adiabatically. Owing to the fact that there is a deep layer of conditionally unstable air above the condensation level, the cumulus clouds will grow upward. Such clouds will have towering heads and protuberances showing internal motion. These are *towering cumulus*, or *cumulus congestus* clouds (see Fig. 25).

Neither the fair-weather cumulus nor the towering cumulus produces precipitation, but the latter may develop into a cumulo-nimbus.

Case 4.—This case (Fig. 49*C*) differs from case 3 only in that the layer of air above the condensation level is conditionally unstable up to such heights that the cloud towers reach up to levels where the temperature is considerably below freezing. The water

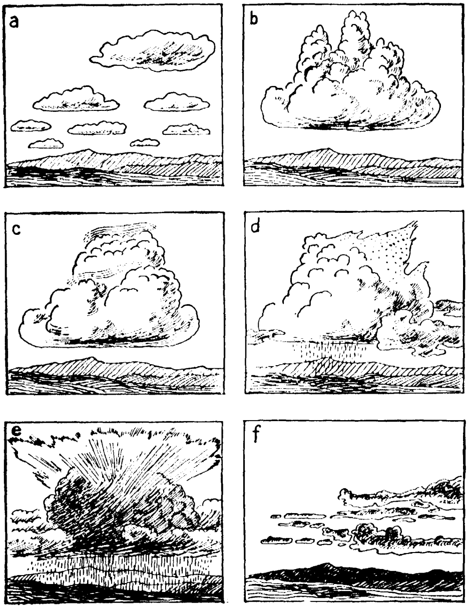


FIG. 50.—Types of convective clouds. *a*, fair-weather cumulus; *b*, towering cumulus; *c*, towering cumulus with "scarf"; *d*, cumulo-nimbus without anvil; *e*, cumulo-nimbus with anvil; *f*, strato-cumulus formed from cumulus.

droplets in the cloud tops begin to freeze. This is an important change in the life history of the cloud, for, as was explained in the chapter on Evaporation, Condensation, and Precipitation, the cloud is then colloidally unstable, with the result that the water droplets accumulate on the ice particles, which then become too heavy to be kept afloat by the ascending currents. Precipitation is then released from the cloud; the towering cumulus has developed into a cumulo-nimbus (shower cloud or thundercloud, see Fig. 27).

Convective Clouds.—The characteristic features of the convective clouds are shown diagrammatically in Fig. 50. The fair-weather cumuli (Fig. 50*a*) are rather flat, without active towers, and well separated from one another. The base is well defined, and their tops show no tendency to grow upward (see also Fig. 24). The presence of fair-weather cumulus indicates that the atmosphere above the condensation level is stable, that the clouds will not develop appreciably, and that the weather is likely to remain fair.

Figure 50*b* shows the typical features of towering cumulus (see also Fig. 25). The presence of protuberances of typical cauliflower structure is characteristic. Internal motion and turbulence are noticeable in the towering cumulus, whereas the fair-weather cumulus has a rather "dead" appearance. Both these types of cumulus consist of water droplets without admixture of ice particles in their tops. However, the towering cumulus may grow to greater heights and develop into cumulo-nimbus.

Figure 50*c* shows a towering cumulus with a fine silky veil, or "scarf," around its top. The scarf usually consists of water droplets; but if the temperature is sufficiently low, it may contain ice crystals. The presence of a scarf often indicates the transition to a cumulo-nimbus. The next stage of the development is illustrated in Fig. 50*d* (see also Fig. 26). The cauliflower structure begins to disappear, and the top of the cloud tends to flatten and develop into a tangled web. This development is presumably due to the formation of ice particles in the upper portion of the cloud. Precipitation is then released. This type of cloud is called cumulo-nimbus calvus, or bald cumulo-nimbus.

We next consider Fig. 50*e*. Here the cumulo-nimbus has grown to formidable proportions and developed an anvillike ice cloud around its top (see also Fig. 27). This type of cloud, which is called cumulo-nimbus incus, or cumulo-nimbus with anvil, usually gives heavy rain squalls, thunder and lightning, and sometimes hail.

It is possible with practice to forecast the weather for a few hours ahead by looking at the cumulus clouds and observing their development. It is then most important to observe the changes in the upper part of the clouds. If there are no towers (as in Fig. 50*a*), there is no chance of precipitation. If there are towers (as in Fig. 50*b*), it is possible that precipitation may develop. If some of the clouds show signs of presence of ice crystals, the precipitation is sure to be released soon. Figure 50*c* often occurs as a transition from Fig. 50*b* to *d*, but Fig. 50*d* may also develop directly from Fig. 50*b*.

The cumulus clouds sometimes dissolve by general shrinking and disappear gradually. This usually occurs when a sheet of high clouds develops above them. In dissolving in this way, they pass through the state of fair-weather cumulus (Fig. 50*a*). Most frequently the cumulus clouds flatten out into rolls or bulging layers; this development is shown in Fig. 50*f*. This often occurs in the evening when the atmosphere is settling down after the diurnal heating. In any case, the dissolution of cumuli shows that the atmosphere is developing toward a stable stratification.

If the convection is caused by diurnal heating over land, it has a pronounced diurnal period with a maximum of cumulus clouds in the afternoon and clearing in the evening. Over oceans the diurnal convection is only slight (except in tropical regions), and the maximum of cloudiness has a tendency to occur in the night.

If the convection is caused by the travel of air toward warmer regions, there is but slight diurnal variation, and cumulus and cumulo-nimbus may develop both day and night.

The precipitation caused by convection is always of a showery character; it begins and ends suddenly, owing to the rapid transition from ascending to descending currents. The sky is variable, with frequent changes from a dark and threatening appearance to clearing.

Thunderstorms.—Thunderstorms develop from clouds of the cumulo-nimbus type in excessively unstable air. The mechanism of their formation may be described briefly as follows:

The falling velocity of raindrops depends on their size. If the drops grow larger than 4 mm. in diameter, they will fall with a velocity exceeding 8 m./sec. When such a high velocity of fall is reached, the drops break up into smaller drops, which then fall less rapidly.

If the ascending currents in the cumulo-nimbus exceed 8 m./sec., the largest raindrops will be split up into smaller drops and will be carried upward. The ascending current in a cumulo-nimbus is not steady; it consists of a succession of gusts and lulls, so that the drops may rise and fall, grow and break up repeatedly.

Each time a drop breaks up into smaller drops, the negative and the positive electricity will be separated, the air taking up a negative charge and the drops a positive charge. By repeated splitting up of drops, enormous electric charges are made available for the thunderstorms. Since the air ascends much more rapidly than the drops that break up, it follows that the positive charge is accumulated in the part of the cloud where the ascending current is strongest and the rest of the cloud becomes negative or neutral.

Severe thunderstorms are often accompanied by hail. The structure of the hailstones shows conclusively the existence of large ascensional velocities in the thunderstorms. The hailstones often consist of concentric shells of clear ice and snow, which shows that the hail must have been moved repeatedly from the liquid to the snow part of the cloud.

A fully developed thunderstorm is accompanied by strong gusts, heavy rain or hail, and lightning and thunder. The wind freshens during the approach of the storm, blowing at first toward the advancing storm. As the thundercloud arrives overhead, the wind changes in direction, blowing out from the storm in a forward direction. The barometer falls while the storm approaches, but when the wind changes a brisk rise amounting to a few millibars occurs. The precipitation, which commenced as a sudden heavy downpour, changes into a more continuous rain which gradually decreases in intensity.

The passage of a thunderstorm is frequently accompanied by strong gusts which may cause complete loss of control of an aircraft.

Figure 51 shows diagrammatically the distribution of electric charges in a thundercloud. The positive charge is counteracted in the core of the ascending current. The rest of the water cloud is usually negatively charged, and the ice cloud carries a positive charge.

Inversions.—The air temperature normally decreases with elevation at a rate of about 0.5°C. per 100 m., or 1°F. per 300 ft. Under special conditions, a reversal of the lapse rate may develop, showing a layer of air in which the temperature increases with height. This is called an *inversion*.

Inversions develop frequently near the earth's surface during calm and clear nights on account of the radiative cooling of the underlying surface (see curve *D*, Fig. 48). They may also develop as a result of cooling when warm air streams over a colder surface. If the wind is sufficiently strong to mix the surface layer, the inversion will be found at some distance above the ground (see curve *E*, Fig. 48). Inversion may also develop in the free atmosphere as a result of large-scale motion of descending air spreading out laterally above the surface layer. The inversion is then caused by the adiabatic heating of the descending air; the air above the inversion is then relatively dry.

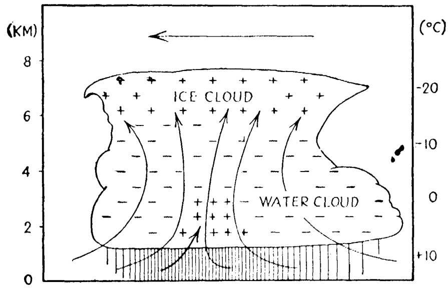


FIG. 51.—Distribution of electric charge in a thundercloud. (*After Simpson.*)

The essential feature of an inversion is the pronounced stability of the air in the layer that has increasing temperature. A well-developed inversion acts as a lid through which no convection or mixing can take place, the ascending currents being repulsed by the excessively stable layer. Since the turbulent mixing tends to transport moisture upward and maintain a steep lapse rate, the relative humidity under the inversion will be high. In pronounced cases, a layer of stratus or strato-cumulus clouds forms immediately under the inversion.

The higher the inversion is above the ground, the deeper is the layer that can be stirred by turbulence. The height of the inversion above the ground has a marked influence on the diurnal range of temperature. The higher the inversion, the deeper is the layer to be heated, and the diurnal variation in temperature at all levels under the inversion will be correspondingly smaller. Conversely, a low inversion favors a large temperature variation. Low inversions are easily destroyed by diurnal heating and mixing of the air, but high inversions usually persist day and night. Turbulence also tends to carry dust and smoke upward; and if an inversion is present, the impurities will not penetrate the inversion. In such cases the haze layer has a distinct upper limit, coinciding with the base of the inversion.

Formation of Fog.—Nonsaturated air may become saturated in three different ways, *viz.:* (1) by evaporation of water into the air, (2) by mixing, and (3) by cooling. When a fog forms, it is necessary that these processes should take place at the earth's surface. In most cases of fog formation the three processes operate together. We shall, however, first discuss each of them separately, and afterward comment on their various combinations.

1. *Evaporation.*—As was shown on page 43, the evaporation is proportional to the factor

$$E_w - e$$

where E_w is the saturation vapor pressure at the water surface and e is the actual vapor pressure of the air. The evaporation will continue until the two vapor pressures are equal, or until

$$E_w = e$$

Let E denote the saturation vapor pressure of the air, which is entirely a function of the temperature, as shown in Fig. 9. Now, if the temperature of the liquid water is lower than that of the air, then

$$E > E_w$$

and the evaporation will continue until

$$E_w = e$$

which means that the evaporation will cease before the air has become saturated.

On the other hand, if the temperature of the liquid water is higher than that of the air, then

$$E_w > E$$

and since the evaporation will continue until $E_w = e$, it follows that balance is reached when

$$e > E$$

which means that the evaporation will tend to continue after the air has become saturated.

In the absence of condensation nuclei, the air would actually become supersaturated; but since nuclei are always present in abundant amounts, the superfluous water will condense, and a fog may result. Therefore, *fogs may form when warm water evaporates into cold air.* This may occur when cold air streams over warm water and also when warm rain falls through cold air. For reasons that will become evident later, we may subdivide these evaporation fogs into two classes, *viz.*: *frontal fogs* and *steam fogs.*

a. Frontal fogs.—As will be shown in a later chapter, a front is a surface of separation between a wedge of cold air and a mass of warm air, as shown in Fig. 52*A*. The warm air, being lighter

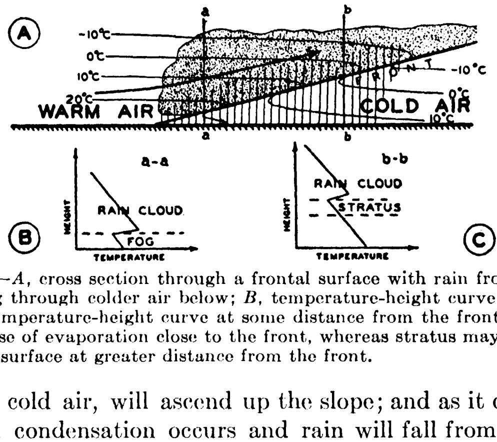


FIG. 52.—*A*, cross section through a frontal surface with rain from warm air aloft falling through colder air below; *B*, temperature-height curve close to the front; *C*, temperature-height curve at some distance from the front. Fog may form because of evaporation close to the front, whereas stratus may form under the frontal surface at greater distance from the front.

than the cold air, will ascend up the slope; and as it cools adiabatically, condensation occurs and rain will fall from the cloud system. The raindrops, coming from the warm air aloft, will be warmer than the air under the front. It will be seen from Fig. 52*A* and *B* that a fog can form only when the frontal surface is close to the ground. Where the frontal surface is high (Fig. 52*C*), the raindrops would first fall through a layer of colder air immediately under the frontal surface and then through a layer of warmer air.

The fogs that form under such conditions will be arranged as a moderately narrow band along the front at the ground. They are therefore called frontal fogs.

The formation of frontal fogs is counteracted by the turbulent mixing. Only when the temperature of the falling rain is much higher than that of the air will frontal fogs form. These fogs, therefore, occur almost exclusively in connection with well-developed fronts.

b. Steam fogs.—Fogs are sometimes observed when cold air streams over a water surface the temperature of which is very much higher than the air temperature. These fogs are called steam fogs or arctic sea smoke. On account of the water being much warmer than the air, the evaporation is so intense that steam pours forth from the water surface and fills the air with fog.

Steam fogs are frequently observed in arctic regions in winter when extremely cold air from adjacent continents streams over open water. Since the air then is heated rapidly from below, it tends to become unstable, and vertical currents will tend to dissipate the fog. However, when a temperature inversion is present at some short distance above the sea, the whole layer of air under the inversion may be filled with fog.

2. *Mixing.*—It was shown in the sections on mixing that the combined effect of horizontal and vertical mixing is to cause the air to become relatively drier near the earth's surface and relatively moister in the upper portion of the mixed layer. Therefore, mixing tends to dissolve fogs and, simultaneously, to cause stratus to form.

3. *Cooling.*—By far the most frequent and most effective cause of fog formation is cooling of the air while in contact with the ground. It is convenient to distinguish among three different causes of cooling, *viz.:* (*a*) radiative cooling of the earth's surface; (*b*) advective cooling, or cooling due to the travel of warmer air over a colder surface; (*c*) adiabatic cooling due to ascending motion of the air. The last cooling is important near the earth's surface only when the air blows up a slope. We may, therefore, speak of *radiation fog, advection fog,* and *upslope fog.*

a. Radiation fog.—The influence of radiative cooling of land areas on the air temperature is of the order of magnitude of 2°F. (1°C.) per hour and is therefore important. The cooling of the air near the earth's surface is greatest on calm and clear nights; for a cloud cover will hinder the outgoing radiation, and strong winds will create turbulence so that the loss of heat is distributed over a deep layer of air. Thus, the conditions favorable for the formation of radiation fog are cloudless sky, calm or almost still air, and high relative humidity. The above conditions occur most frequently in continental areas when there is descending motion aloft so that the clouds dissolve. As high relative humidity is an important factor, radiation fogs develop most often in air of maritime origin when it becomes stagnant over cold continents.

b. Advection fog.—The rate of cooling of traveling air masses depends on the difference in temperature between the air and the underlying surface. In extreme cases, this cooling may amount to 4°F. (2°C.) per hour. As high wind velocity favors vertical mixing, it follows that advection fogs cannot readily form in strong winds. On the other hand, very slight winds are unfavorable for the production of advection fogs, for the rate of cooling of the air will then be small. As a result, it is found that advection fogs occur most frequently with moderate wind velocities, say 5 to 15 m.p.h. Advection fogs with wind velocity above 30 m.p.h. are exceedingly rare.

Advection fogs occur fairly often over land in winter when air from the oceans invades the cold continents. Since the radiative temperature changes are relatively large over continents, most land fogs are caused by the combined influence of advective and radiative cooling.

At sea, the diurnal variation of the temperature of the underlying surface is exceedingly small (usually less than 1°F.). The sea fogs are therefore either advection fogs or frontal fogs. The sea fogs occur most frequently over the cold ocean currents (*e.g.*, the Labrador Current) during invasions of warm air. They also occur quite frequently along the Gulf coast in winter on account of the cold water transported southward by the Mississippi. The summer fog along the California coast is, to a certain extent, caused by advection of warm air over the cold upwelling water along the coast.

c. Upslope fog.—When air blows up a slope, it cools adiabatically; and if the relative humidity is sufficiently high, it will reach its condensation level, and a fog may form. Such fogs, which are called upslope fogs, very often form on the eastern slope of the Rocky Mountains when warm moist air streams up the slope.

The formation of upslope fogs is counteracted by turbulence, as is the formation of all kinds of fog. Moreover, in order to obtain fog through upslope motion, it is necessary that the air, when saturated, be stably stratified, for otherwise convectional currents would develop. Therefore, upslope fogs can form only when the air is convectively stable before the ascent commences. When this is the case, upslope fogs may or may not form, depending on the intensity of turbulence.

Diurnal Variation of Fog.—All types of fog have a tendency to dissipate through heating; there is, therefore, a marked diurnal variation in the frequency of fogs, with a maximum in the early morning hours and a minimum in the afternoon. Naturally, a shallow fog burns off more readily than a deep fog. Since upslope fogs and advection fogs are usually deeper than radiation fogs, the latter type has a more distinct diurnal variation than the former.

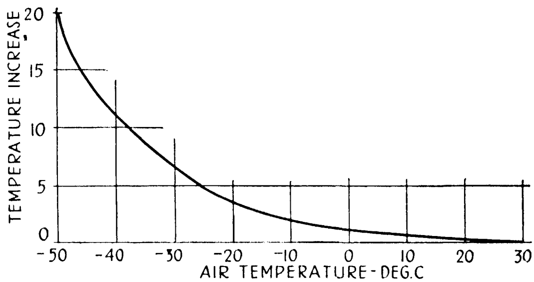


FIG. 53.—Temperature increase necessary to accommodate the water vapor resulting from the evaporation of a fog containing 0.3 gram of liquid water per cubic meter.

Other conditions being equal, the diurnal variation is the more pronounced the higher the temperature. A dense fog contains about 0.3 gram of liquid water per cubic meter. In order to dissipate the fog, this water must be evaporated; in addition, the air temperature must be increased so much that this additional amount of water vapor can be accommodated in the air. It follows, then, from Fig. 9 that only a slight increase in air temperature suffices to accommodate the fog water when the air temperature is high but that a considerable amount of temperature increase is required when the air temperature is low. A simple computation gives the values contained in Fig. 53. Furthermore, it is well known that the diurnal variation in temperature is large when the air temperature is high and small when the air temperature is low. For both reasons, fogs that occur at high temperatures (in the warm season) will have a distinct tendency to burn off in the early morning, whereas fogs that occur at low temperatures (in the cold season) do not readily dissolve because of the diurnal heating. Fogs that occur in tropical climates, therefore, have a very regular daily rhythm, whereas those occurring in arctic and polar continental climates in winter have only a slight diurnal variation.

Fog over Snow-covered Ground.—It was mentioned on page 46 that a mixture of water droplets and ice particles is colloidally unstable, with the result that the water droplets tend to accumulate on the ice. This effect vanishes at 0°C., and it increases with decreasing temperature below freezing. It follows then that a water fog over snow-covered ground will tend to dissolve when the temperature is below freezing.

When the snow is melting, the air temperature above the snow is above freezing, and the temperature of the air that is in contact with the snow is 0°C. There is then a temperature inversion at the ground, and the eddy transfer (see page 71) of heat will be directed downward towards the snowy surface. The heat thus carried downward is consumed in the melting of the snow.

Under normal conditions, the specific humidity of the air increases with elevation when there is a temperature inversion above the snow. The eddy transfer of moisture will then be directed downward, and water vapor will condense on the snow. Thus, when snow is melting, heat and moisture are carried toward the snowy surface, and the process tends to cause the air to become drier. It follows, then, that fogs do not readily form over melting snow when the air temperature is much above freezing.

To sum up, we may say that melting snow has a marked tendency to dissipate fogs. This dissipating influence increases the higher the air temperature is above 0°C. As the air temperature decreases to 0°C., this dissipating influence vanishes; but when the air temperature falls below freezing, another dissipating influence, due to the depression of the saturation pressure over ice (see Fig. 32), commences and increases in intensity until the air temperature reaches -10 to -15°C., where it has a maximum. It follows then that a water fog is in a stable state over snow only when the air temperature is not far removed from 0°C.

The above conclusions are corroborated by observational evidence. An example is shown in Fig. 54. It should be noted that October is a snowless month, January and February are snow-covered, with temperatures mostly below freezing, and March is characterized by melting snow. It will be seen from Fig. 54 that in October the fog frequency increases steadily with decreasing air temperature, as one would expect over snowless ground. When snow is present on the ground, the frequency curve has a totally different trend. The maximum fog frequency occurs, not at the lowest temperature, but when the air temperature is close to freezing.

Frequency curves similar to Fig. 54 for stations in north Siberia show that the fog frequency over snow increases when the air

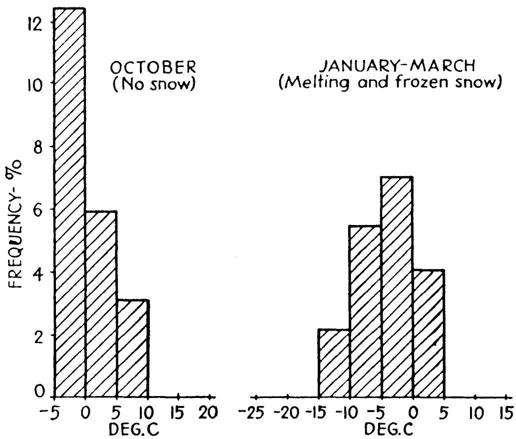


Fig. 54.—Fog frequencies. Oslo, Norway, 1920–1931.

temperature sinks below $-35°C$. This is due to the fact that the fogs that occur at such low temperatures are mostly ice-crystal fogs, and these are in equilibrium with respect to the underlying surface.

Classification of Fog.—To sum up the foregoing discussion, the fogs may be classified according to the identity of the process that constitutes the principal cause of their formation. A classification based on this principle is given in the table on page 92. A summary of the fog-dissipating agencies appears in the right-hand portion of the table.

Distribution of Fog.—Since sea fogs are mostly of the advection type, the ocean areas noted for high frequency of fogs coincide with the cold ocean currents.

The Labrador Current and the Oya Shio have frequent fogs throughout the year, with a maximum in summer. In winter, it is mostly the warm maritime air from the south that produces fog in these waters; in summer, fogs are frequently found when warm air from the adjacent continents is cooled over the cold water.

CLASSIFICATION OF FOG

Fog-producing Processes	Fog-dissipating Processes
I. Evaporation from:	I. Sublimation or condensation on:
1. Rain that is warmer than the air (rain-area fog, or frontal fog)	1. Snow with air temperature below 0°C. (excepting ice-crystal fog)
2. Water surface that is warmer than the air (steam fog)	2. Snow with air temperature above 0°C. (melting snow)
II. Cooling due to:	II. Heating due to:
1. Adiabatic upslope motion (upslope fog)	1. Adiabatic downslope motion
2. Radiation from the underlying surface (radiation fog)	2. Radiation absorbed by the fog or by the underlying surface
3. Advection of warmer air over a colder surface (advection fog)	3. Advection of colder air over a warmer surface
III. Mixing:	III. Mixing:
1. Horizontal mixing (unimportant by itself and strongly counteracted by vertical mixing)	1. Vertical mixing (important in dissipating of fogs and producing stratus)

The cold equatorward ocean currents off the coast of California and north Africa (from Casablanca to Senegal) have many fogs in summer. Frequent fogs are also observed along the west coast of South Africa from 8 to 32°S. (June to August), along the east coast of Africa south of Cape Guardafui (June to September), along the east coast of South America south of Rio de la Plata (July to September), and along the west coast of South America from 4 to 31°S. (August and September). These fogs are formed because of the cooling of the air by the passage over the cold coastal currents.

The Norwegian Sea and the Barents Sea have frequent fogs in summer which form through advection of warm air from the adjacent continents. The same also applies to the North Sea and the Baltic, but the fogs occur here most frequently in spring or early summer when the sea surface is cold, partly on account of

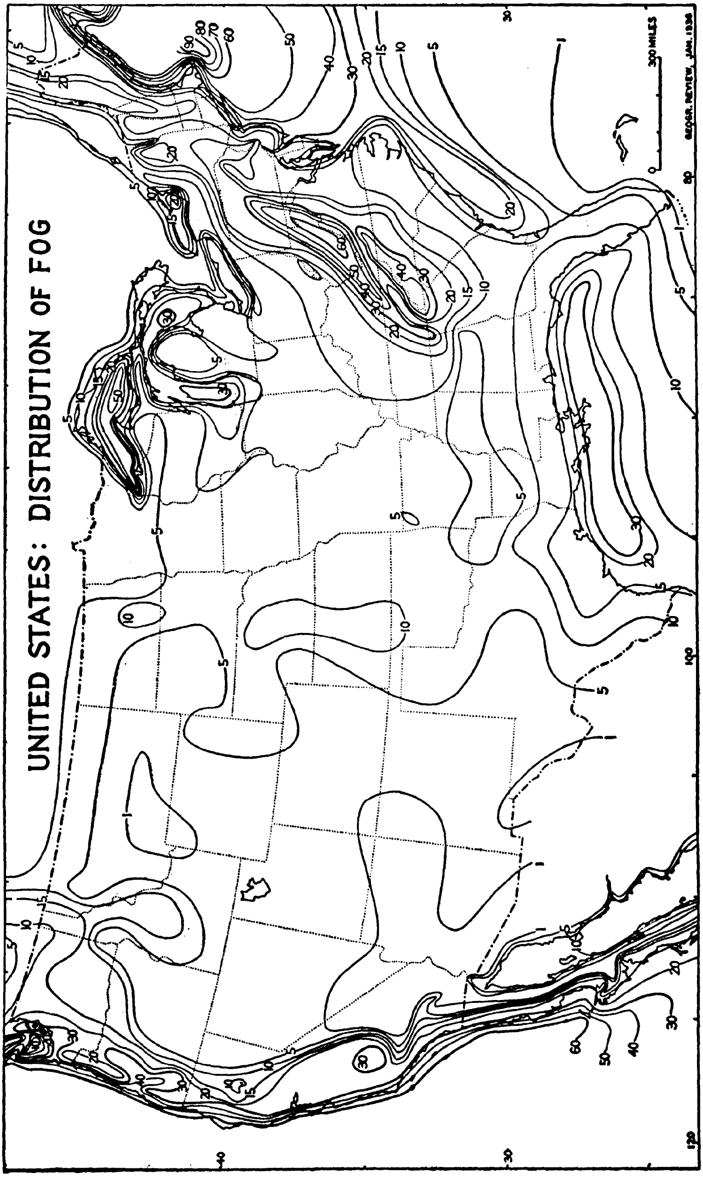


FIG. 55.—Map showing the number of days of dense fog (with visibility less than 1000 ft.) in the United States.

the ice water from northern continents, and warm air is produced over central Europe and other warm sources.

A detailed study of the distribution of fogs in the United States has been made by R. G. Stone, the results of which are summarized in Fig. 55 and in Table VI.

TABLE VI.—FOG IN THE UNITED STATES

Area	Season of maximum frequency	Type of fog
Appalachian Mountains	Late summer (September)	Mixed types: radiation, upslope, advection
California coast	Summer, secondary maximum in February	Advection
Great Lakes	Spring	Advection
	Autumn	Steam, radiation, frontal
Middle Atlantic coast	Early spring (February, March)	Advection
New England coast	Summer	Advection
	Winter	Frontal
Northern Pacific coast	Summer	Advection
Pacific coast valleys	Winter	Radiation
Gulf coast	Winter (December and January)	Advection

Ice Accretion.—Icing is one of the greatest dangers to air navigation, and its formation cannot as yet be forecast with sufficient accuracy, chiefly on account of lack of adequate observations from the free atmosphere.

Ice usually forms on the forward edges of wings and struts or on the propeller, but sometimes it forms also on the horizontal faces. The ice that forms on the forward edges of the wings changes the profile of the airfoil. Often, the cross section of the ice deposit is of a mushroom shape. This causes a general change in the streamlines around the wings, with the result that the aircraft may lose so much in dynamic lift that flying becomes disastrous.

Ice that forms on the propeller is dangerous because of the irregular rotation that results from the asymmetrical distribution of the weight of the ice. There is no force perpendicular to the axis of the propeller in the case of perfect symmetry; but when ice gathers unevenly or when a lump of ice breaks off from one blade of the propeller and not from the other, a force proportional to the *square* of angular velocity of the propeller acts perpendicularly to the axis of the propeller. This may cause destructive vibrations.

Icing may also occur in the carburetor, and controls may freeze. Such icing may be eliminated by the aid of heating arrangements.

Modern deicing equipment has proved capable of overcoming icing of slight and moderate intensity on the propellers and on the forward edges of the wings, but no adequate means has as yet been invented of eliminating intense icing, or icing that takes place on the horizontal faces. Icing, therefore, remains a potential danger to air navigation.

It is known that icing may occur (1) outside the clouds, (2) within the clouds, and (3) in subcooled rain that falls from warmer air aloft into a colder layer of air.

As far as temperature range is concerned, it is known that icing may occur on aircraft from about 36 to about −20°F. (+2° to −30°C.), although the critical temperature interval seems to be from 32 to 14°F. (0 to −10°C.).

1. *Icing in Cloudless Air.*—In the same way as hoarfrost forms on the ground through the process of sublimation, ice deposits may form on the aircraft in cloudless air below freezing. The intensity of such icing is slight, and the ice deposits are so crisp that they break off easily.

2. *Icing within Subcooled Water Clouds.*—The icing that occurs within the clouds varies greatly in intensity. In extreme cases, it may amount to ¼ in./min. or more, and sometimes no icing occurs at all.

To understand the icing process, it is necessary to bear in mind the following facts:

a. When a subcooled drop freezes, its temperature rises to 0°C.

b. To raise the temperature of 1 gram of water 1°C., 1 gram calorie is needed (the specific heat of water).

c. When 1 gram of water freezes, 80 gram calories are liberated (the latent heat of fusion).

d. To evaporate 1 gram of water, about 590 gram calories are required (the latent heat of vaporization).

In explaining the accretion of ice on an object, it is necessary to account for the removal of the latent heat of fusion. The processes through which the latent heat of fusion can be removed are (1) adiabatic cooling due to the pressure variation along the airfoil, (2) conduction of heat to the air, (3) conduction of heat to the object, (4) heat convection, and (5) evaporation of water. Computations as well as experiments have shown that the last factor, *viz.*, the evaporation of water, is the most important.

To illustrate the process of freezing, we consider 1 gram of subcooled water whose temperature is $-10°C$. On striking the airfoil, freezing commences and the temperature of the freezing water rises to 0°C. This accounts for 10 gram calories; but there remain still 70 gram calories to be removed in order to freeze 1 gram of water. Since the latent heat of vaporization is approximately 590, it follows that the evaporation of 0.12 gram of water suffices to remove the remaining 70 gram calories. Thus, only a minute amount of evaporation is required to remove the necessary heat. Since the saturation vapor pressure increases with the air temperature (see Fig. 9), it follows that the vapor pressure over freezing water is higher than the vapor pressure in a subcooled cloud. The freezing water will evaporate, and this will cause the drop to solidify.

3. *Icing with Temperatures above Freezing.*—If the temperature of the cloud is slightly above freezing, the water on the aircraft must first be cooled to freezing and then the latent heat of fusion removed in order that ice may form. This occurs very rarely; but when it occurs, the removal of the heat is mainly due to evaporation of water.

4. *Icing in Subcooled Rain.*—The physical process is exactly the same as when icing occurs in a subcooled cloud. However, since the raindrops are much larger than the almost microscopical cloud droplets, the ice accretion in subcooled rain will be more intense. Furthermore, the falling raindrops will strike the horizontal as well as the vertical faces, whereas the ice that comes from subcooled clouds will accumulate only on the forward edges. Thus, subcooled rain is the most dangerous icing condition.

Subcooled rain occurs when rain falls from warmer air aloft into a layer of colder air with temperatures below freezing.

5. *Deposition of Drops.*—The smallest cloud droplets tend to follow the air motion around the wing. It happens, therefore, quite frequently that no ice forms on the aircraft within a subcooled cloud. The larger the droplets, the greater is their inertia; as a result, the large drops will strike the aircraft more readily

than do the small drops. Another important factor to consider is the number of drops per unit volume. Obviously, other conditions being the same, more ice will form in a dense cloud than in a thin cloud. Thus, the most favorable conditions for icing are large cloud droplets and large content of liquid water (*i.e.*, numerous droplets). The following table gives computed values of icing intensity in inches per minute for various drop sizes and liquid-water contents.

COMPUTED ICING INTENSITIES (IN./MIN.) WITH AIR SPEED ABOUT 100 M.P.H.

Drop diameter, mm.	Amount of liquid water	
	5 grams/cu. m.	0.5 gram/cu. m.
0.006	1/32	Nil
0.012	3/16	1/64
0.018	5/16	1/32
0.024	3/8	1/32
0.032	1/2	1/16

The size of the drops that can be suspended in a cloud depends on the ascending velocity. Since the ascensional velocity is largest in convection clouds, it follows that the icing intensity is likely to increase with the lapse rate within the cloud.

6. *Rate of Freezing.*—Since the freezing of water must proceed at the rate with which the latent heat of fusion is removed, it follows that a small drop will freeze faster than a large drop. The smallest cloud droplets solidify almost instantaneously on striking the object, whereas the larger cloud droplets and the raindrops freeze more slowly. Now, if the rate of freezing is slow, the next drop will be deposited before the previous drop has solidified. In such cases, the deposit takes the form of glaze (solid, amorphous ice). On the other hand, if the droplets are so small that they solidify completely before other drops are deposited, the ice deposit takes the form of rime, with a porous and granular structure.

CHAPTER VII

WIND SYSTEMS

As far as motion is concerned, the atmosphere may be regarded as an engine that creates kinetic energy (wind) from heat energy, the difference in temperature between the poles and the equator and between the upper and lower atmosphere being the significant sources of energy.

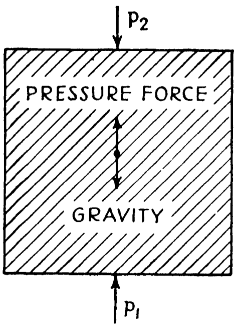


FIG. 56.

Since the atmospheric pressure varies in the horizontal as well as in the vertical direction, a unit of air will be exposed to a force that tends to move it from high to low pressure. The motion thus created would be modified by friction. Furthermore, when the motion is referred to the rotating earth, a force, due to the rotation, will have to be considered. Finally, the force of gravity controls the motion along the vertical. Before we discuss the wind systems, we shall comment on each of these four forces.

The Pressure Force.—The forces that move the air depend primarily on the distribution of pressure. Let us first consider a vertical cross section through a cube of air with horizontal and vertical faces (Fig. 56). Since the atmospheric pressure decreases with elevation, the pressure p_1 on the lower face of the cube is greater than that of p_2 on the top face. This represents a force that is directed upward. This force is counteracted by the weight of air within the cube, or the gravity force. Usually, there is balance between the two forces, so that no vertical motion results. The equation of static equilibrium (page 51) then holds.

Occasionally, there is not complete balance between these forces, and vertical accelerations result. In this way, convective currents are created. The mean vertical velocity over large areas is, however, exceedingly small, and rarely exceeds 3 in.

(7 cm.)/sec. The large wind systems are therefore mainly horizontal currents.

The pressure also varies in the horizontal direction. The pressure on one of the vertical faces of the cube in Fig. 56 will then exceed the pressure on the opposite face. The difference in pressure is equivalent to a force tending to drive the cube horizontally in the direction from high to low pressure. Since there is no component of the gravity force in the horizontal direction, a horizontal pressure force, acting on resting air, will be unbalanced, and it will create and maintain motion. The motion thus created will be modified by friction and the rotation of the earth.

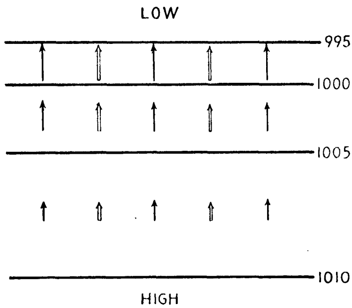


FIG. 57.—Relation between isobars and pressure gradient (single-shaft arrows). If the pressure force were the only force acting on the air, the wind would blow as indicated by the double-shaft arrows.

Pressure Gradient and Isobars.—Suppose that we observe the atmospheric pressure in a large number of places in a horizontal surface (*e.g.*, sea level) and plot the pressures on a map. Let us next draw curves through the points that have identical pressures. Such curves are called *isobars*. They are usually drawn for every fifth millibar,[1] as shown in Fig. 57. *The horizontal pressure gradient in a point may be defined as the decrease in pressure per unit distance in the direction in which the pressure decreases most rapidly.* It will be seen from Fig. 57 that the pressure gradient, as represented by the single-shaft arrows, (1) is everywhere perpendicular to the isobars, (2) points from high to low values of pressure, and (3) is inversely proportional to

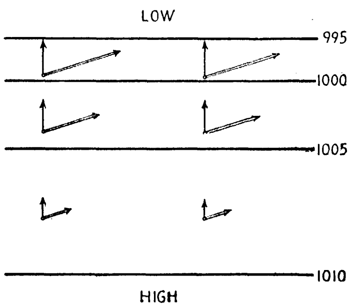


FIG. 58.—Isobars, pressure gradient (single-shaft arrows), and wind (double-shaft arrows) as actually observed near the earth's surface.

[1] In the United States, isobars are often drawn for every third millibar.

the distance between the isobars: the more crowded the isobars, the stronger is the pressure gradient.

Now, if the pressure force was the only force acting on the air, the air would move in the direction of the pressure gradient as shown by the double-shaft arrows. This distribution of wind velocity does *not* agree with experience. What is actually observed in nature is shown in Fig. 58 where the wind blows mainly along the isobars, but with a slight drift toward lower pressure. From this we may conclude that the pressure force is not the only force that acts on the air. We shall therefore consider the influence of the earth's rotation.

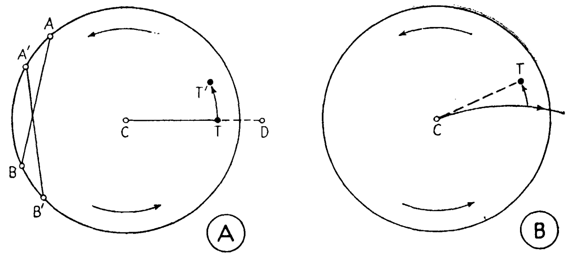


FIG. 59.—Motion relative to a rotating disk.

The Deflection Force.—In considering the relation between the forces that act on the air and the motion of the atmosphere, it is necessary to bear in mind that the observed winds represent the motion of the air relative to the rotating earth.

The influence of the rotation on the relative motion may be demonstrated in the following manner: We consider a circular disk that rotates around the center C as shown by the arrows in Fig. 59A. A bullet is shot from the center C of the disk at a target T that partakes in the rotation. At the moment the bullet is shot, the line from the center to the target runs through the point D which, being outside the disk, does not partake in the rotation. The law of inertia requires that the bullet moves along the straight line from C to D; but while the bullet does so, the target T will rotate from T to T'. An observer on the rotating disk, not seeing the point D, would obtain the impression that the bullet moved somewhere to the right of the target T, whereas another observer outside the disk, not seeing the target T, would think that the bullet moved on the straight line CD. Both observers would be right in their statements; the one observes the *relative* motion, and the other observes what we may call the *absolute* motion.

Since we are here concerned with the relative motion, it is of interest to consider the path of the bullet relative to the rotating disk. At the initial moment, the bullet would move in the direction of the moving target T. However, as the target rotates to the left, the bullet will curve off to the right, and its path, referred to the rotating disk, would be a curved line as shown in Fig. 59B.

In a similar manner, it may be shown that the rotation of the disk gives an apparent deflection to the right when the bullet is shot from the edge of the disk toward the center, and also when it is shot in either direction perpendicular to the radius of the disk. Thus, for example, if a bullet is shot from A to B, the observer at A and the target at B would rotate from A to A' and from B to B', respectively. Thus, while the bullet moves on the straight line from A to B, the observer arrives at A' and sees the target in the direction $A'B'$, and, again, the path of the bullet would appear to be deflected to the right of the target.

In considering the motion relative to a rotating disk, the deviation to the right of the "line of force" may be accounted for by a force that produces an acceleration equal to

$$2V\omega$$

where V is the relative velocity of the moving particle and ω is the angular velocity of the rotating disk. This force, which is called the *deflecting force*, or the *Coriolis force*, would everywhere be at right angles to the path and to the right of the direction of the motion.[1]

The above example of a rotating disk illustrates the conditions at the North Pole, but a similar effect is present also in lower latitudes. To show this, we consider Fig. 60A, which represents a cross section through a meridian circle. We consider now the motion relative to the earth in an arbitrary point A whose latitude is φ. We place a cone tangent to the earth's surface along the parallel circle AB. This cone unfolded is pictured in Fig. 60B, it represents the horizon plane in latitude φ. Now, while the

[1] The reason why bullets, after all, hit targets on the rotating earth is that their path is so short that the deflection is small and also that the guns are so constructed that the deflection is reduced to a minimum.

earth rotates the angle 2π in 24 hr., the point A on the unfolded cone rotates the angle $2\pi \sin \varphi$. Thus, if ω is the angular velocity of the earth, the point A on the unfolded cone would have an

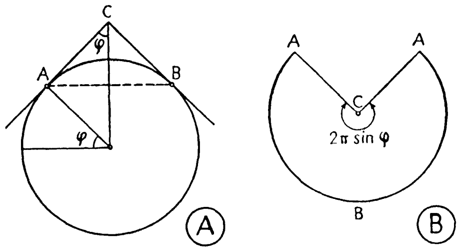


FIG. 60.—*A*, Meridional cross section through the earth with cone tangent to latitude φ; *B*, the cone unfolded, representing the horizon plane in latitude φ.

angular velocity equal to $\omega \sin \varphi$, and the deflecting force which is $2V\omega$ at the pole would be

$$2V\omega \sin \varphi$$

in the latitude φ.

It will be seen that the deflecting force has its maximum value at the pole ($\sin \varphi = 1$), and it decreases to zero at the equator; below the equator, $\sin \varphi$ is negative, and the deflecting force is to the left of the direction of the relative motion.

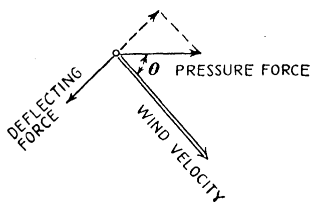


FIG. 61.

The reason why the deflecting force in the horizontal direction varies with latitude is that the centrifugal force due to the earth's rotation is tangent to the earth's surface at the poles and perpendicular to the earth's surface at the equator.

The Geostrophic Wind.—If a unit of air, originally at rest, is exposed to a horizontal pressure force, it will start moving in the direction of the pressure gradient. However, because of the deflecting force due to the earth's rotation, it will be accelerated to the right of the direction of the pressure gradient, and, after a while, its velocity will be as shown in Fig. 61. It will be seen that part of the pressure force is balanced by the deflecting force. However, the pressure force has a component along the velocity, with the result that the motion will be accelerated. Since the deflecting force tends to make the moving unit deviate farther from the pressure force, the angle θ in Fig. 61 will increase and the wind velocity will increase until the motion is at right angles to the pressure force. This is shown in Fig. 62. Since the pressure force is perpendicular to the isobars, it follows that the motion will eventually become parallel to the isobars with low pressure to the left and high pressure to the right of the direction in which the air moves.

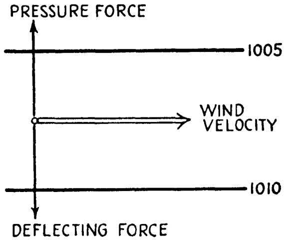


FIG. 62.—Balance between the pressure force and the deflecting force.

It will be seen from Fig. 62 that the pressure force is balanced by the deflecting force. The motion is then steady: there is no acceleration. Thus, if a unit of air is allowed to move under the influence of the pressure force and the deflecting force alone (no friction), the motion would eventually become parallel to the isobars. We should then obtain a picture as shown in Fig. 63. Comparing this with Fig. 58, we see that the pressure force and the deflecting force cannot account for the slight drift across the isobars that is actually observed in nature. From this we may conclude that there must be a third force (friction) that modifies the motion.

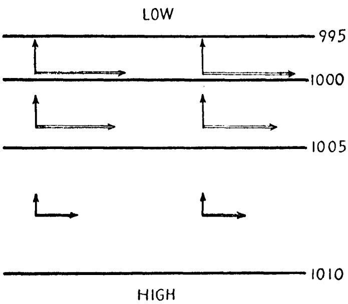


FIG. 63.—Relation between wind and isobars when the pressure force is balanced by the deflecting force. The wind resulting from this balance is called the geostrophic wind.

Before we go on to discuss the influence of friction, we shall consider in greater detail the balance between the pressure force and the deflecting force. We have already seen that the acceleration due to the deflecting force is

$$2V\omega \sin \varphi$$

The deflecting force acting on a unit cube of air of density ρ must then be equal to

$$\rho 2V\omega \sin \varphi$$

because

$$\text{Force} = \text{mass} \times \text{acceleration}$$

Furthermore, the pressure force acting on a unit cube of air is simply equal to the pressure gradient, which is denoted by G. Since the deflecting force is balanced by the pressure gradient, we obtain

$$G = \rho 2V\omega \sin \varphi$$

Solving with respect to V, we obtain

$$V = \frac{G}{\rho 2\omega \sin \varphi}$$

This formula gives us the wind velocity that would result if the pressure gradient is completely balanced by the deflecting force. This is called the *geostrophic wind.*

Let Δp denote the pressure difference over a certain distance H. Then

$$G = \frac{\Delta p}{H}$$

Now, the pressure difference between the isobars in Fig. 63 is 5 mb., or 0.5 centibar.[1] Therefore,

$$G = \frac{0.5}{H} \text{ centibars}$$

which substituted into the equation for the geostrophic wind gives

$$V = \frac{0.5}{\rho 2\omega \sin \varphi H}$$

Since the air density is almost constant in the horizontal direction, it follows that *the geostrophic wind is inversely proportional to the distance between the isobars.* Moreover, the geostrophic wind blows along the isobars with high pressure to the right and low pressure to the left of the direction of the motion (as shown in Fig. 63).

[1] Centibar is the appropriate unit for pressure when length is expressed in meters, density in metric tons per cubic meter, and time in seconds. These are the units commonly used in theoretical meteorology.

In the Southern Hemisphere, the deflecting force is to the left of the motion. Therefore, in the Southern Hemisphere, the geostrophic wind blows along the isobars with high pressure to the left and low pressure to the right of the direction of the motion.

The above rules may also be expressed as follows: If you stand with the back toward the wind, then you have *low* pressure on your *left*-hand side and *high* pressure on your *right*-hand side. In the Southern Hemisphere, the reverse is true.

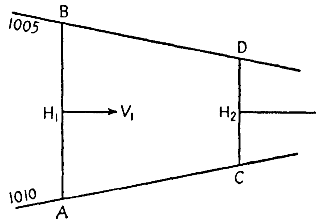


FIG. 64.—The inflow through AB equals the outflow through CD when the wind is geostrophic.

Let us next consider the flow between two isobars as shown in Fig. 64. The flow through the cross section AB is

$$V_1H_1$$

and the flow through the cross section CD is

$$V_2H_2$$

The net outflow D from the area $ABDC$ is then

$$D = V_2H_2 - V_1H_1$$

Applying the formula for the geostrophic wind to H_2 and H_1 and substituting in the above formula, we obtain

$$D = 0$$

which means that there is no inflow (convergence) to or outflow (divergence) from the area $ABDC$. This, again, means that the amount of air that streams out of the section CD is exactly equal to the amount that streams in through the section AB. We may, therefore, say that *the geostrophic wind has no divergence or convergence.* The geostrophic wind blows between the isobars in the same manner as the water streams in a river: the narrower the cross section of the river, the faster is the flow. We may therefore think of the space between the isobars as *tubes of constant transport of air.*

It will be shown presently that the actual wind is very close to the geostrophic wind whenever the frictional influence is slight.

The Influence of Friction.—It is easy to prove that friction modifies the flow in such a manner that the air tends to flow across the isobars in the direction from high to low pressure, the drift across the isobars that is actually observed (see Fig. 58) being thus accounted for.

Suppose that the air moves with the velocity **V** as shown in Fig. 65. The deflecting force **D**, we know, is at right angles to the velocity, and the frictional force **F** is opposite to the direction of **V**. The resultant of the deflecting force and the frictional force is indicated by the arrow **R** in Fig. 65. Now, to obtain balanced motion, the pressure force **G** must be equal to the resultant **R** and have a direction opposite to that of **R**. Since the angle from **V** to **R** is greater than 90°, it follows that the angle from **V** to **G** must be less than 90°. Furthermore, since the pressure force is perpendicular to the isobars, it follows that the wind **V** must blow to the left of the isobars, as shown in Fig. 65.

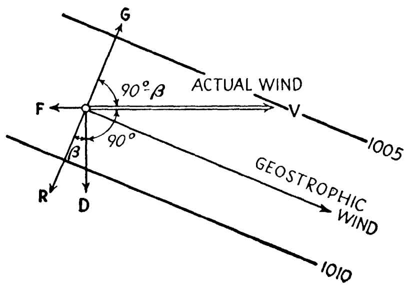


FIG. 65.—Showing the relation between wind and isobars when the pressure force is balanced by the resultant of the deflecting force and the frictional force. When the frictional force decreases, the angle β decreases, and the wind blows more along the isobar.

The angle between the isobar and the wind direction increases with increasing friction; it is therefore greater over land than at sea. Under normal conditions, this angle is 25 to 35°. Thus, the actual wind near the earth's surface is deviated to the left of the direction of the geostrophic wind (the direction of the isobars).

As we ascend from the earth's surface, the frictional force diminishes, and we approach the state characterized by the balance between the deflecting force and the pressure force as shown in Fig. 62. Observations from the free atmosphere show that the frictional force is negligible at a distance of 3000 ft. above the ground. For all practical purposes, we may say that the actual wind between 2000 and 4000 ft. conforms with the geostrophic wind as indicated by the isobars at sea level. In this way *the isobars at the ground are indicative of the wind velocity and direction at intermediate levels.*

Wind Variation with Height.—Since the frictional force decreases with elevation, the wind velocity will increase as we ascend through the atmosphere. Simultaneously, the wind direction will change so as to become more parallel to the isobars. Figure 66 shows the normal distribution of wind velocity with height, and Fig. 67 shows how the wind direction varies under normal conditions.

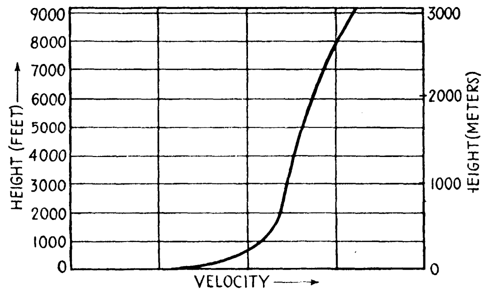


Fig. 66.—Normal variation in wind velocity with height.

The layer below 3000 ft., within which the influence of friction is noticeable, is called the *friction layer.* It will be seen that the wind increases most rapidly near the ground and approaches the geostrophic wind gradually. At the surface, the actual wind over land is, on the average, about 40 per cent of the geostrophic wind, whereas at sea it is about 70 per cent. The difference is due to the reduced friction over water surfaces.

Observations from the free atmosphere show that the wind velocity under normal conditions continues to increase with elevation above the friction layer. This increase, which is almost linear, continues up to the tropopause, above which the velocity again decreases. The increase in wind above the friction layer is due to the fact that the pressure force varies with elevation.

Types of Pressure System.—If the atmospheric pressure is observed at a large number of stations over a large area and if isobars are drawn, it will be found that only a limited number of types of pressure distribution occurs. Figure 68 shows diagrammatically the various types of isobaric picture and the corresponding winds in the Northern Hemisphere.

The principal types of pressure distribution are *areas of high and low pressure.* The areas of low pressure are usually called *depressions, cyclones,* or simply *lows.*

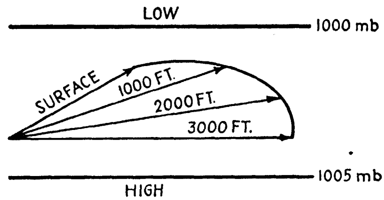


FIG. 67.—Normal variation in wind direction within the friction layer.

A depression, or cyclone, may then be defined as an area within which the pressure is low relative to the surroundings. It will be seen from Fig. 68 that the wind circulation around a cyclone (low) is counterclockwise, with a slight drift toward the left of the isobars.

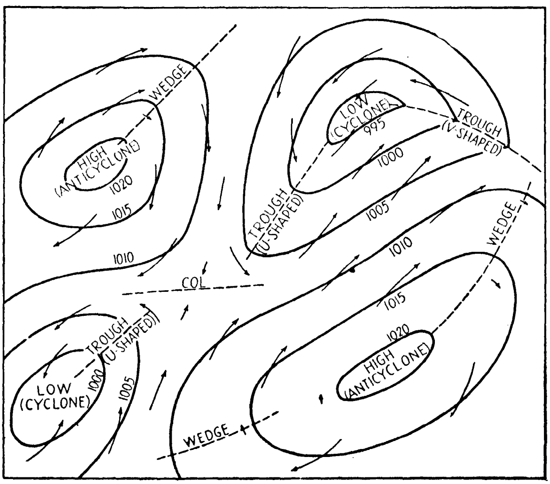


FIG. 68.—Types of pressure systems.

An anticyclone may be defined as an area within which the pressure is high relative to the surroundings. The wind circulation is clockwise around an anticyclone, with a drift away from the center.

A *trough* of low pressure is an elongated area of relatively low pressure which extends from the center of a cyclone. The trough may have U-shaped or V-shaped isobars (see Fig. 68), the latter being associated with fronts.[1] The wind circulation around a trough is essentially of the cyclonic type.

A *wedge* of high pressure is an elongated area of high pressure that extends from the center of an anticyclone. The wind circulation is essentially anticyclonic.

A *col* is the saddle-backed region between two anticyclones and two cyclones arranged as shown in Fig. 68.

In all pressure systems, the wind blows mainly along the isobars with a slight drift toward lower pressure. Thus, the winds in an anticyclone blow spirally out from the center, and those in a cyclone blow spirally toward the center. In the very center, the

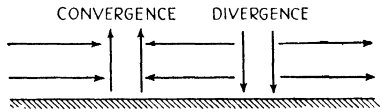


FIG. 69.—Ascending and descending currents resulting from convergence and divergence, respectively.

pressure gradient vanishes, and either there is calm or there are light and variable winds. The same is true of the col.

The winds indicated in Fig. 68 refer to the conditions near the ground. About 3000 ft. (or 1000 m.) above the ground, the wind would blow almost along the isobars.

Divergence and Convergence.—Although the geostrophic wind has no convergence or divergence, the same is not true of the actual wind. The drift of air across the isobars causes inflow of air toward the cyclonic regions and outflow of air from the anticyclonic regions (see Fig. 68). Since there can be no permanent piling up of air in the cyclones or permanent loss of air in the anticyclones, the air that converges must ascend, and the air that diverges must descend (see Fig. 69). Since the ascending air cools adiabatically and the descending (subsiding) air heats adiabatically, the cyclonic regions will, in the main, be characterized by clouds and precipitation, whereas the anticyclonic regions will, in general, be regions of fair weather. Convergence and divergence in the horizontal flow may also result from other causes. The above rules need, therefore, not hold in each individual case, but they apply to the majority of cases.

[1] To be discussed later.

The General Circulation.—An idealized diagram showing the principal features of the pressure distribution and the prevailing winds is shown in Fig. 70. Around the equator, there is a region of almost uniform pressure, in which the winds are light and variable. This belt of light and variable winds is called the *doldrums.* It will be seen that the winds converge from both hemispheres into the doldrums. This convergence results in ascending air currents, adiabatic cooling, condensation, and

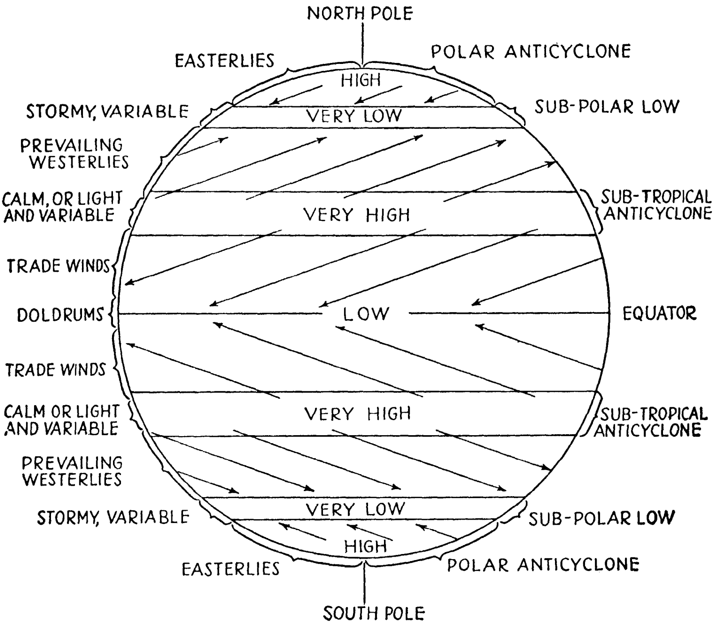


FIG. 70.—Idealized model of the pressure distribution and the prevailing winds at the surface of a uniform earth.

precipitation. The doldrums are, therefore, characterized by frequent showers, thunderstorms, and heavy rainfall. The doldrums are generally found slightly to the north of the equator; they have a slight annual oscillation, tending to move toward the summer Hemisphere.

Farther away from the equator are belts of high pressure with easterly winds on their equatorial sides and westerly winds on their poleward sides. These belts of high pressure are called the *subtropical anticyclones.* The winds on their equatorial sides are called the *trade winds.* They blow mainly from the east and have a component toward the equator; on the poleward sides, the winds have a poleward component. The subtropical anticyclones are regions of divergence and descending (subsiding) air currents, low relative humidity, almost clear sky, and deficit of rainfall. Most deserts are found within this region. The subtropical anticyclones move slightly poleward in summer and toward the equator in winter.

In the central portion of the subtropical anticyclones, the winds are light and variable. This belt is often referred to by seamen as

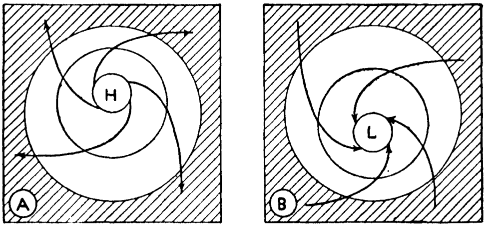


FIG. 71.—Hatched areas represent oceans, unhatched areas represent continents. In winter (*A*), the pressure increases over the cold continents; and in summer (*B*), it decreases. This gives rise to monsoon winds.

the *horse latitudes.* The winds on the poleward side of the high-pressure belts are called the *prevailing westerlies.* They increase in strength as the latitude increases.

The prevailing westerlies are bounded on their poleward sides by a region of low pressure in subpolar latitudes (the *subpolar low*), on the polar side of which there are easterly winds around the anticyclones that occupy the ice-covered polar regions.

The Monsoons.—The wind and pressure distribution shown diagrammatically in Fig. 70 corresponds to what would be observed if the earth's surface were uniform. Because of the difference in temperature between continents and oceans, this zonal arrangement is considerably modified, particularly in the Northern Hemisphere. In the Southern Hemisphere, the continental areas are relatively small, with the result that the distribution of pressure and winds is approximately as shown in Fig. 70. In the Northern Hemisphere, the annual heating and cooling of the large land areas cause pronounced annual variations, with winds blowing from sea to land in summer and from land to sea in winter. Such winds are called *monsoons*, although the word usually refers to the monsoons in south Asia only.

The mechanism of the monsoon winds may be described briefly as follows: Let Fig. 71 represent an idealized continent in middle and high latitudes where the annual variation in temperature is considerable. In winter (Fig. 71*A*), the air over the continent

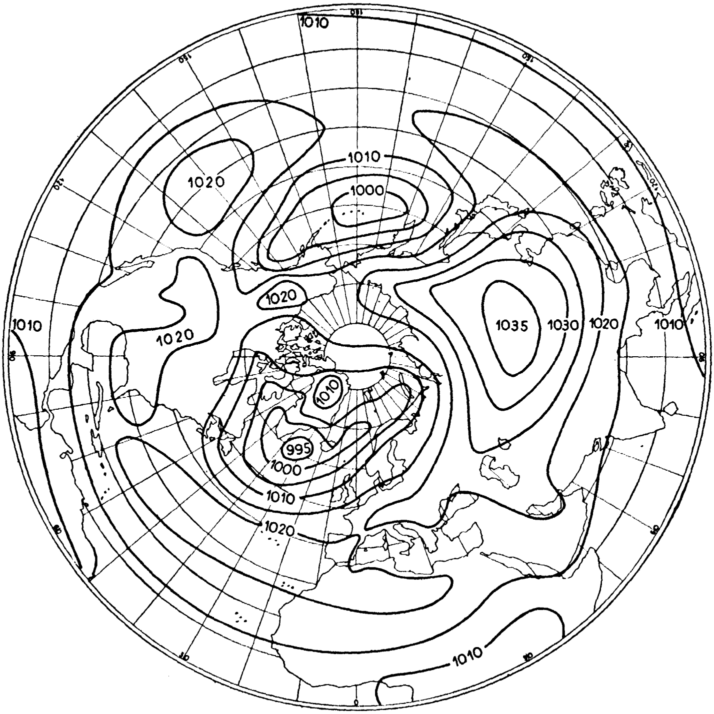


FIG. 72.—Mean distribution of pressure at sea level in winter.

cools and becomes denser than the air over the adjacent oceans. Since the atmospheric pressure is equal to the weight of the air column, it follows that the pressure will increase over the cold continents as the winter approaches, and a cold anticyclone will tend to form. This will cause a wind circulation as shown in Fig. 71*A*.

As the summer approaches, the anticyclone dies away and is replaced by a low-pressure area over the continents, which are now warmer than the adjacent oceans. The pressure distribution and the wind circulation will then be as shown in Fig. 71*B*.

Now, the *actual* pressure distribution and wind circulation in the Northern Hemisphere will then be as follows:

FIG. 73.—Surface air currents in midwinter.

In winter. A zonal circulation consisting of uniform girdles as shown in Fig. 70 with a circulation like Fig. 71*A* superimposed on the continents.

In summer. A zonal circulation consisting of uniform girdles as shown in Fig. 70 with a circulation like Fig. 71*B* superimposed on the continents.

Maps for the Northern Hemisphere showing the mean pressure distribution and the wind circulation at the earth's surface are seen in Figs. 72 and 73 for winter conditions, and in Figs. 74 and 75 for summer conditions.

It will be seen from Fig. 72 that cold anticyclones are present over the continents in winter. As a result, the subpolar low (see Fig. 70) is disrupted and is present over the northern parts of the Pacific and Atlantic oceans. These lows are more or less stationary and permanent; the one over the northern part of the Atlantic Ocean is called the *Icelandic Low,* and the one over the

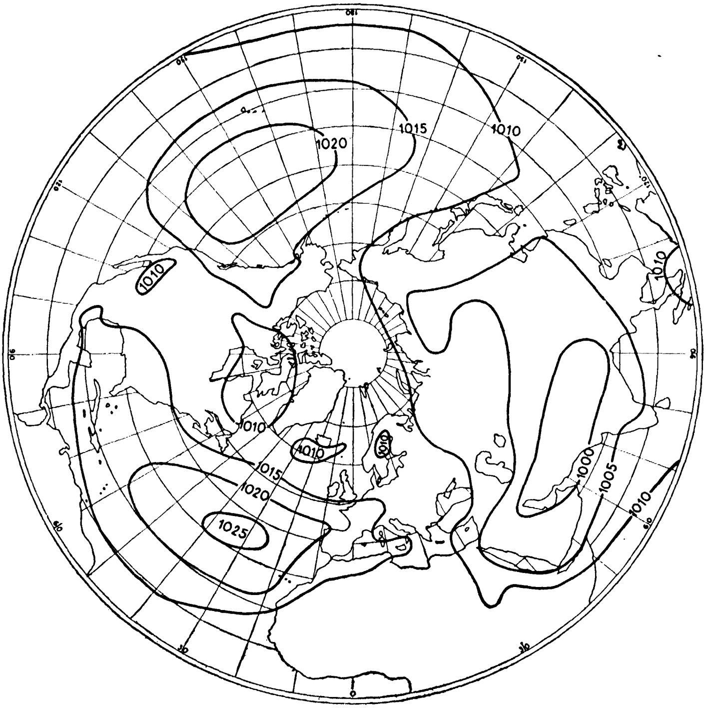


FIG. 74.—Mean distribution of pressure at sea level in summer.

northern part of the Pacific is called the *Aleutian Low.* The high-pressure area over Asia in winter is usually called the *Siberian High,* and the one over the United States is called the *North American High.* These highs and lows together with the subtropical anticyclones over the oceans are the principal *centers of action.*

In summer (see Fig. 74) the subpolar lows decrease in intensity and the oceanic highs increase in intensity.

It will be seen that, on the whole, the observed pressure distribution in winter conforms with what we should obtain if we superimposed Fig. 71*A* on Fig. 70, and that in summer with what we should obtain if we superimposed Fig. 71*B* on Fig. 70. Thus the general circulation of the lower layers of the atmosphere

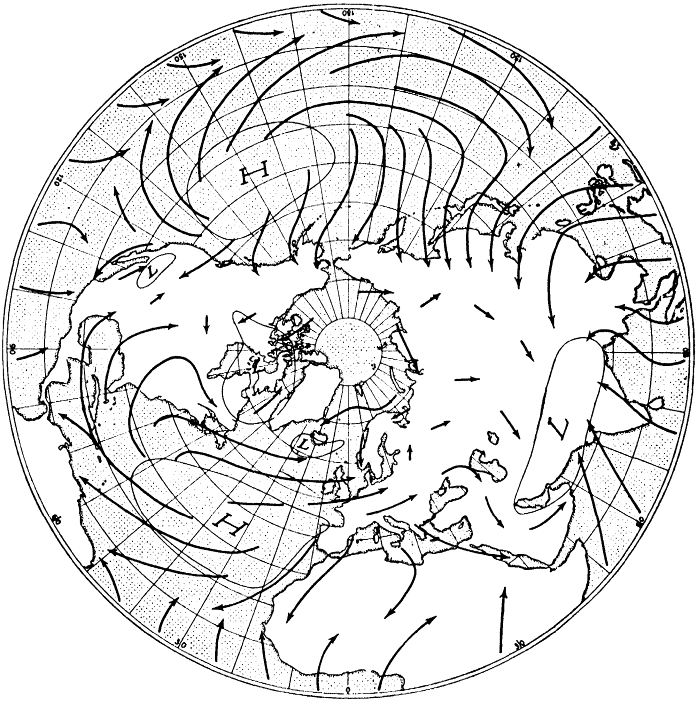


FIG. 75.—Surface air currents in summer.

consists partly of a purely zonal circulation and partly of a monsoon circulation.

It will be seen that the monsoon system is most pronounced in Asia where land areas are most extensive. A monsoon system is present also over the North American continent, but it is weaker and less persistent than that in Asia. In spring and autumn, the temperature difference between ocean and continents is small. The general circulation is then more like the zonal systems shown in Fig. 70.

Land and Sea Breezes.—In addition to the seasonal variation of temperature and pressure over land and water, there is a daily contrast that exercises a similar but more local effect. In summer the land is warmer than the sea by day and colder than the sea by night. The slight variations in pressure thus established cause a system of breezes with a component landward during the day-

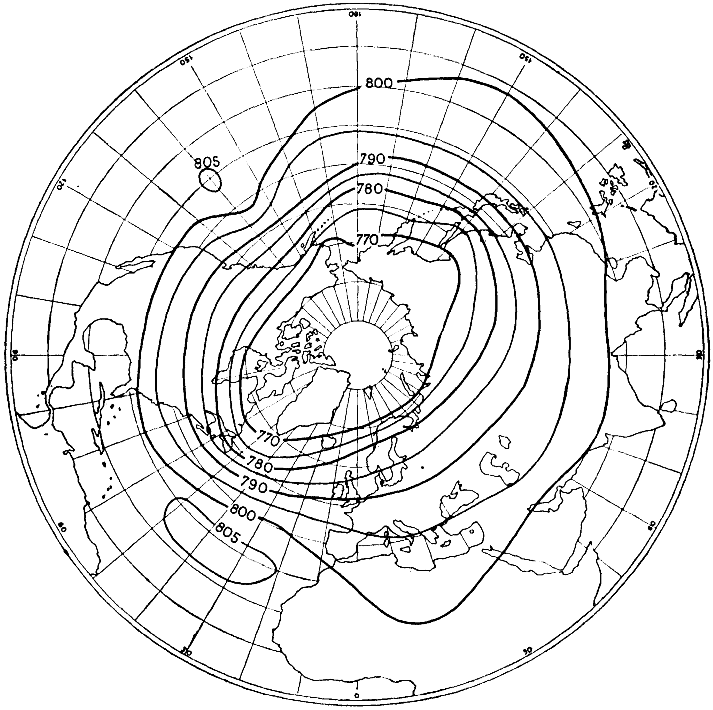


FIG. 76.—Mean distribution of pressure at the 2-kilometer level in winter.

time and seaward during the night. These breezes are shallow and do not penetrate far inland. The day breeze may attain the strength of a fresh or a strong breeze, but the night breeze is usually gentle.

In the tropics the land and sea breezes blow with great persistence. In higher latitudes, they are often overshadowed by stronger winds of more general character.

Mountain and Valley Winds.—On warm days, it is often observed that the winds tend to blow up the slopes during the day and down the slopes at night. This is due to the fact that the mountain slopes are warmer than the free atmosphere at the same level during the daytime and colder than the free atmosphere during the nighttime. Since the cold air tends to sink and the warm air tends to rise, a system of local mountain winds develops with an upslope motion during the day and a downslope motion during the night. These winds do not normally attain appreciable velocities, although the cold downslope winds in winter may occasionally become quite strong, particularly when they blow through narrow valleys.

The Circulation of the Free Atmosphere.—The disturbing influence of the oceans and continents decreases rapidly with elevation, and above the 6000-ft. level the distribution of land and sea has practically no influence on the general circulation, except in India, where the monsoon disturbance is still noticeable.

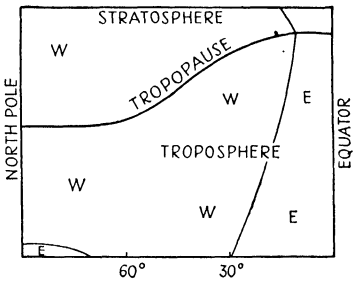


FIG. 77.—Distribution of westerly winds (*W*) and easterly winds (*E*) in the free atmosphere.

The winds above the 6000-ft. level are controlled by an immense low-pressure area centered over the North Pole (see Fig. 76), a belt of high pressure in the subtropics, and a belt of uniform pressure around the equator. As a result of this pressure distribution, the mean wind in the free atmosphere is westerly everywhere on the poleward side of the subtropical high-pressure belt. The westerly winds increase with altitude and attain their maximum speed at the tropopause. On the equatorial side of the subtropical high-pressure belts the mean wind blows from an easterly direction. Figure 77 shows diagrammatically the distribution of westerly and easterly winds in a meridional cross section from the equator to the North Pole.

The various circulations described above represent the *mean* state of motion of the atmosphere. On this mean state of motion are superimposed many disturbances which account for the variability of the weather. We shall discuss these in a later chapter.

Turbulence.—The main source of turbulence in the atmosphere is the friction along the surface of the earth. The roughness of the surface creates eddies or whirls of air which are forced up to higher levels. Records of the details of the wind structure show that the turbulent flow consists of a succession of gusts and lulls, the period of which is irregular and of the length of a few seconds. The strength of the gusts is somewhat proportional to the roughness of the ground and the velocity of the wind, and it increases as the stability of the air decreases.

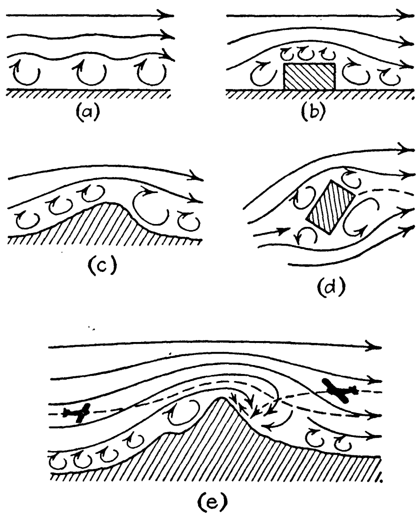


FIG. 78.—Eddies.

Turbulence caused by friction along a rough surface is called *mechanical turbulence.* It is usually limited to a layer of air of about 3000 ft. in thickness. This turbulent layer along the earth's surface may be visualized from Fig. 78, which shows eddies along the surface of the earth and a fairly steady flow above.

Another source of turbulence in the atmosphere is the irregular distribution of temperature. The warmer lumps of air will rise and the colder will sink, irregular flow being thus produced. This kind of turbulence may be called *thermal turbulence.* The mechanical and the thermal turbulence combine to produce the resultant turbulence.

The thickness of the turbulent layer along the earth's surface depends mainly on the stability of the air. The mechanical turbulence is damped in stable air and intensified in unstable air. The thickness of the turbulent layer is, therefore, greater in unstable air than in stable air.

In unstable air the convectional currents are superimposed on the general flow. They cause fluctuations in the wind similar to those characteristic of pure turbulence, but with periods ranging from a few minutes to an hour or more. Such wind variations are called *wind squalls* if they are strong. The wind squalls are usually connected with convective clouds, and the approach of a towering cumulus or a cumulo-nimbus may be taken as a warning of the arrival of wind squall.

Although the pure turbulence is most pronounced near the surface of the earth, the vertical wind squalls are strongest in the free atmosphere, notably in the convective clouds and in their immediate vicinity. A skilled pilot will be able to estimate the chances of severe gusts and squalls from the appearance of the convective clouds.

The intensity of the thermal turbulence and the convectional gusts and squalls depends largely on the nature of the surface. The amount of heat absorbed by the earth's surface depends to a large extent on its moisture content and on its color. The midday temperature on sunny summer days is higher over macadam and sandy fields than over grassland and is lower over wet ground and woods and still lower over water. These differences in temperature give impulses of considerable intensity to local convectional currents. Local currents of this kind are the main cause of the bumpiness experienced above level country on warm days. The convective currents also form easily on the sunward slopes of hills and mountains. In hilly country, therefore, turbulence, local eddies, and convection combine to make the air very bumpy.

Turbulence and Obstacles.—Figure 78*a* illustrates the formation of eddies in the air flow above level ground. If such a current meets an obstacle, the distribution of the turbulence is greatly modified, the resulting turbulence depending on the dimensions and shape of the obstacle, the speed of the current, and the stability of the air. We can here discuss only a few types of obstacles and their influence on the air current.

Figure 78*b* shows the type of eddies that form when an air current passes an obstacle of rectangular cross section. A *stationary* eddy forms on the windward side. Small irregular traveling eddies develop above the roof of the obstacle, re-forming continuously along the windward edge. On the lee side, eddies form and travel down-wind. The lee eddy is usually stationary while it develops; when it has attained a certain intensity, it leaves the obstacle and travels with the wind while it dissipates gradually.

If the horizontal dimensions of the obstacle are small, the air current has a tendency to stream around it. In this case, eddies with vertical axes form at the edges, as shown in Fig. 78*d*. Whether the air will stream round the obstacle or across it depends mainly on the length of the obstacle and the stability of the air. The resulting turbulence caused by large buildings, hangars, etc., is usually a combination of horizontal and vertical eddies.

Figure 78*c* shows the eddies that develop when the current crosses a small ridge. If the incline on the windward side is not too steep, the stationary eddy disappears and there are only the usual eddies caused by the roughness of the surface. On the lee side, there are usually larger eddies that form on the slope and travel down-wind. If the wind is strong and the air unstable, landing may be unpleasant or even unsafe on the lee side close to the ridge.

The Influence of Mountain Ranges.—The influence of mountain ranges on air currents is, in general, the same as that of the obstacles discussed in the previous paragraph. Figure 78*e* shows a cross section through a mountain range. The most striking feature is the well-developed eddy on the lee side of the mountain. On the windward side, there may be a stationary eddy or not, according to whether the incline is steep or not.

The eddies that form on the lee side are often dangerous to air navigation. A *pilot flying against the wind may, if he does not keep sufficient altitude, be forced down onto the mountainside, or he may lose control of the plane in the eddy.* A pilot flying with the wind will usually gain altitude while he approaches the mountain range. If the mountain is very steep, there may be a stationary eddy on the windward side, which may cause difficulties. Under these conditions, the pilot must carefully consider the effect of ascending and descending air currents in making turns or banks with a heavy load.

The eddies around mountain ranges reach up to some altitude above the range, causing intense mixing of the air. With a favorable distribution of humidity and temperature, this mixing will lead to the formation of clouds (stratus) around and above the mountain range. The process of formation of stratus has been described in a previous chapter (see page 70). What we are here concerned to emphasize is that stratus has a tendency to form in mountainous country owing to the increased turbulence (see Fig. 83*H*). A general layer of stratus will be lower on the mountainsides than in the free air. Often, stratus forms only on the mountainsides and not in the free air.

Apart from forming eddies and local clouds, the mountain ranges affect the streamlines of the general flow on a large scale. At a considerable altitude the streamlines are unaffected by the mountain range, and the current is mainly horizontal. It follows, then, that the cross section of the current is diminished by the range. The speed of the current, therefore, will increase in proportion and attain a maximum velocity above the range. The general flow has an upward component on the windward side and a downward component on the lee side. This large-scale influence on the main current is noticeable at great distances. The downward component of the general flow dissolves clouds, and this effect often reaches 200 miles or more to the lee. The ascending current is a frequent cause of general cloudiness and precipitation on the windward slope (see Fig. 83*F* and *G*). If the air is stable, it flows more smoothly and has a tendency to stream around mountain ranges. Unstable air, on the other hand, streams easily across, and the upward movement favors the formation of convective clouds and showers.

Bumpiness.—Most frequently, bumpiness is caused by upward or downward currents in the air. These currents may be caused by turbulence, convective currents, or eddies caused by obstacles. The bump is felt when the aircraft flies into or out of a rising or descending current.

Another cause of bumpiness is sudden horizontal variations in the wind. If the aircraft flies with the wind, a sudden lull will cause a sudden increase in dynamic lift, which is felt as an upward bump. Likewise, a sudden horizontal gust will cause an abrupt decrease in dynamic lift, which is felt as a downward bump. If the aircraft flies against the wind, a lull would give a downward jolt, and a gust an upward jolt.

Bumpiness is also experienced when the aircraft passes a temperature inversion (see page 83). Temperature inversions are usually wind discontinuities, and both the velocity and the direction of the wind are different above and below the inversion. The following example suffices to demonstrate the cause of this kind of bumpiness: An aircraft flies under the inversion with the wind. Its air speed is v, and its ground speed is V, V being greater than v. If the air above the inversion is calm, the aircraft will arrive there with an air speed equal to V. This sudden increase in the air speed gives increased dynamic lift, which is felt as an upward bump. If the aircraft flies under the inversion against the wind and passes through the inversion into calm air above, it will experience a sudden loss in dynamic lift and will drop down again under the inversion. Under the control of the elevator the plane may rise, only to receive another downward bump or a series of bumps. If the aircraft changes its direction sufficiently, it may pass through the inversion with hardly any bumping.

It happens sometimes that ripples or billows develop along the inversion. The aircraft, flying horizontally close to the inversion, may then be exposed to a series of bumps in a regular period. These can be avoided by a slight change in altitude.

Temperature inversions have a very pronounced effect on lighter-than-air craft flying through them. An airship flying in the air above the inversion will have acquired a temperature close to that of the surrounding air. When diving down through the inversion, it dives into colder air. The superheat (*i.e.*, excess temperature) of the ship may then be sufficient to overcompensate the rudder control, and the ship rises above the inversion. Likewise, if an airship flies in the cold air under the inversion and attempts to rise through it, it will receive a downward jolt, for the gas of the ship is colder than the air above the inversion. The difficulties are particularly great if the inversion is close to the ground, for the downward jolt received by the ship when entering the inversion may be sufficient to bump the ship against the ground.

CHAPTER VIII

AIR MASSES

The ideas underlying the principles of the modern methods of weather analysis and forecasting are based on the fact that the general circulation of the atmosphere has a tendency to produce vast masses of air whose physical properties are more or less uniform within large areas, the transition from one such mass to another being rather abrupt. Thus, the large semipermanent anticyclones in the general circulation are either predominantly continental or predominantly oceanic. Each of these anticyclones constitutes a "wheel" in the circulation, and the air within each of these wheels, by circulating over a more or less uniform surface, tends to become uniform as far as temperature and moisture are concerned.

Life History of Air Masses.—Air that remains in contact with the earth's surface will gradually acquire properties characteristic of the underlying surface. If the underlying surface is uniform and if the air currents are favorable, the air mass will tend to become uniform in the horizontal direction. An air mass may therefore be defined as *a huge body of air whose physical properties, notably temperature and humidity, are more or less uniform horizontally.* The regions where such air masses are formed are called *air-mass sources.*

When an air mass leaves its source, it begins to change its physical properties, and the weather phenomena that develop within the mass depend entirely on its *life history.* In studying the life history of an air mass, the following factors should be considered:

1. *The source* whence the air obtained its fundamental properties. Depending on the nature of the source, whether dry or moist, warm or cold, the air mass may be dry or moist, stable or unstable.

2. *The path* that the air mass followed since it left its source. The air mass will change its properties and structure en route from the source to the destination according to whether it travels toward colder or warmer, moister or drier regions (see **Fig. 48**).

3. *The age* of the air mass, or the time it has spent on its journey from its source. The amount of change in the air-mass properties depends largely on the nature of the surface over which it travels and on how long it has been in contact with the surface. The absorption of properties commences at the surface and proceeds upward because of turbulent mixing. The maximum of modifying influence is, therefore, always near the earth's surface when the air mass is "young." How far upward the influence penetrates depends entirely on the nature of the influence and the age of the mass.

The transition from one air mass to another is usually rather abrupt. It is along the border between adjacent air masses that

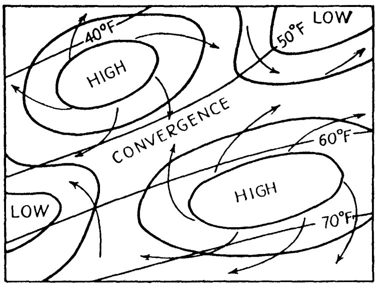


FIG. 79.—If the isotherms move with the air currents, the air within the anticyclones will tend to become uniform while temperature contrasts will be established in the belt of convergence.

the greatest contrasts in energy are found, and it is there that the cyclones (depressions) develop. These phenomena will be discussed in a later chapter. What we are here concerned with is the weather phenomena that develop *within* the air masses. For this purpose, it is necessary to classify the air masses according to the principal types of source regions and developments.

Air-mass Sources.—Since uniformity is the essential feature of an air mass, we may look for the principal source regions in those parts of the world where the earth's surface is reasonably uniform and where, simultaneously, the large-scale air currents are divergent.

The influence of divergent and convergent air currents is illustrated diagrammatically in Fig. 79. It will be seen that the temperature lines (isotherms) in the divergent wind systems will tend to move apart, the air tending to become more uniform.

The opposite is the case where the air currents converge. Here, the isotherms will be brought together, and temperature contrasts will be established.

It is then readily understood that the most favorable condition for effective production of uniform air masses is an anticyclonic wind system centered in a region where the underlying surface is

FIG. 80.—Air-mass sources of the Northern Hemisphere in winter. 1, arctic; 2, polar continental; 3, polar maritime, or transitional; 4, transitional; 5, transitional, or tropical maritime; 6, tropical continental; 7, tropical maritime; 8, equatorial; 9, monsoon.

reasonably uniform. On this principle, we may classify the air-mass sources as follows:

1. *The Arctic and Polar Continental Source.*—In winter, this source occupies the arctic fields of snow and ice and the snow-covered portions of North America and the Eurasian continent (see Fig. 80). This region is, on the whole, swept by anticyclonic

winds. In summer (Fig. 81), only the northernmost portion of these continents can be classified as sources for polar continental air masses.

2. *The tropical maritime sources* occupy the regions of the subtropical anticyclones. The underlying surface is uniform, and the wind system is predominantly anticyclonic. The principal

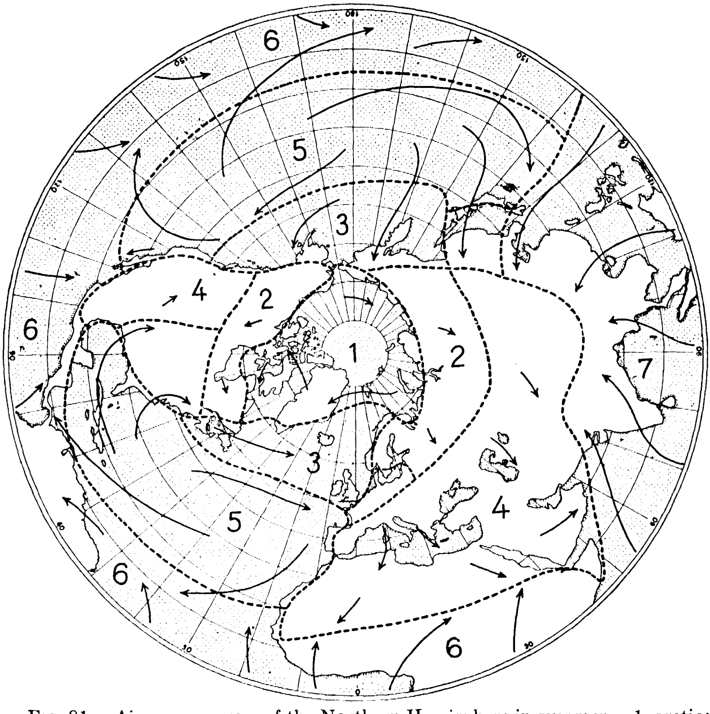


FIG. 81.—Air-mass sources of the Northern Hemisphere in summer. 1, arctic; 2, polar continental; 3, polar maritime; 4, tropical continental; 5, tropical maritime; 6, equatorial; 7, monsoon.

regions are indicated in Figs. 80 and 81, for winter and summer conditions.

3. *The tropical continental source* is, in winter, limited to the North African continent and in summer to the vast expanses in Africa, Asia, and Southern Europe. In North America, tropical continental air is produced during the summer in the arid regions west of the Mississippi.

4. *The polar maritime source* is found in the north and northeastern part of the Atlantic and Pacific oceans (see Figs. 80 and 81). In winter, there is a steady flow of air from the polar continental sources toward the adjacent oceans. Within the transitional regions indicated by 4 in Fig. 80, the polar continental air is transformed into polar maritime air, and within the regions indicated by 5 the polar continental air is transformed into tropical maritime air.

5. *The equatorial source* occupies the equatorial belt between the trade winds. The underlying surface is extremely uniform, with hardly any annual variation. Since this belt is predominantly oceanic, the equatorial air has a high moisture content; and since the underlying surface is warm, it is usually unstable.

6. *The monsoon source* occupies the southern and southeastern part of Asia. In winter (Fig. 80) the cold air from the polar continental source streams persistently across the mountain ranges in Asia toward the equator. In summer (Fig. 81), air from the equatorial region streams into the interior of Southern Asia. As a result, the winter season is relatively cold and extremely dry, whereas the summer season is characterized by high temperature, high relative humidity, and excessive rainfall.

Classification of Air Masses.—The idea of air masses was first introduced into meteorological literature and forecasting practice by Bergeron, who introduced a dual classification of air masses, *viz.*,

1. A geographical classification based on the principal source regions described in the foregoing section.

2. A thermodynamical classification based on the influences sustained by the air masses when they travel out of the typical source regions.

The geographical classification, as it is now commonly used, is summarized in Table VII.

The thermodynamical classification comprises two main types, *viz.:*

a. Cold air masses, or air that is colder than the underlying surface. An air mass of this type will absorb heat from below and will change its lapse rate as indicated by the curves A, B, and C in Fig. 48.

b. Warm air masses, or air that is warmer than the surface over which it travels. An air mass of this type will give off heat to the underlying surface; it will be cooled from below, and its lapse rate will change as indicated by the curves A, D, and E in Fig. 48.

Whereas the geographical classification refers rather to the *integral* result of the processes that lead to the formation of more or less uniform air masses as a result of prolonged sojourn in the principal source regions, the thermodynamical classification is

TABLE VII.—GEOGRAPHICAL (SOURCE) CLASSIFICATION OF AIR MASSES

Symbol	Denomination	Winter source Fig. 80	Summer source Fig. 81	Remarks
A	Arctic	1	1	Unimportant in midsummer
Pc	Polar continental	2	2	Pronounced in winter
Pm	Polar maritime	3	3	
Tc	Tropical continental	6	4	Prominent in summer
Tm	Tropical maritime	7 and 5	5	
E	Equatorial	8	6	
M	Monsoon	9	7	
S	"Supérieur"	Formed in the free atmosphere through descending motion in anticyclones in middle latitudes		

essentially *differential* in character and refers mainly to the recent developments that have led to the formation of clouds, precipitation, and other passing phenomena within the air masses.

The classifications described above supplement one another. Thus, the term "a cold mass of maritime polar air" indicates a vast body of air of polar origin which is traveling over an ocean that is warmer than the air itself. Taking into consideration the season and the time that this air mass has been under maritime influence, the above term is descriptive of a definite type of weather phenomena characteristic of the mass as a whole.

The Properties of a Cold Mass.—The source of the cold masses is normally in the subpolar or arctic regions; but, in winter, masses of cold air may develop over continents down to 30°N. While in their sources, these masses are cooled from below and are characterized by

1. Stable stratification, notably in the lower layers.
2. Low specific humidity.
3. Low temperature.

When such a mass, for some reason or other, moves toward warmer regions, it will arrive there with a temperature lower than that of the surface over which it travels. The mass will be heated from below, and thermal instability will soon develop in the lower layers and gradually spread upward. If the air originally contained inversions, these will be destroyed by continued heating from below, with the result that a uniform steep lapse rate develops throughout the mass; this results in convective currents.

If the cold mass travels over water, it will pick up moisture, which is brought up to higher and higher levels by the convective currents. Convective clouds form and soon develop into cumulonimbus. If the cold mass travels over land, it will be heated from below, but it will not absorb much moisture. In this case, therefore, convective clouds do not easily form until the instability has reached up to very great altitudes. Continental cold masses are, therefore, often accompanied by fair weather.

The following examples will show the outstanding difference between a cold mass that travels over a continent and one that travels over an ocean: (1) Summer air from, say, Saskatchewan streams southward toward Texas. In spite of being heated 30°F. or more, it remains clear or perhaps produces a few scattered cumuli, because it does not absorb much moisture while it is heated. (2) Cold and dry winter air from, say, Texas streams across the Gulf of Mexico whereby it may be heated 15°F. At a distance of about 100 miles off the coast, towering cumulus and cumulo-nimbus develop, because the air has been heated and has absorbed moisture from below.

A typical *maritime* cold mass is recognized by the presence of several or all of the following characteristics:

1. Increasing temperature and humidity.
2. Steep lapse rate and instability.
3. Turbulence, gusts, and squalls.
4. Cumulus and cumulo-nimbus.
5. Variable sky, changing from dark and threatening to bright or clearing.
6. Showery precipitation that commences and ceases abruptly.
7. Good visibility between showers because marine polar air is fairly pure in its source and has not picked up dust on its journey.
8. Height of the base of the clouds moderate, but rarely less than 1000 ft. This condition is due to the heating of the air from

below whereby the condensation level is kept at some distance above the ground.

A typical *continental* cold mass is recognized by the presence of several or all of the following characteristics:

1. Increasing temperature and fairly constant humidity.
2. Steep lapse rate and instability.
3. Scattered cumulus clouds, and occasionally cumulo-nimbus.
4. Pronounced diurnal period in cloudiness with a maximum in the afternoon.
5. Precipitation, if occurring, of a showery character with considerable bright intervals.
6. Visibility variable but on the whole good, except when the air has traveled for a fairly long time over dusty land or industrial regions.
7. Height of the base of the clouds considerable and rarely less than 2000 ft.

When a maritime cold mass invades a continent in summer, the instability is intensified; the showers increase in both frequency and intensity. When a maritime cold mass invades a continent in winter, the instability decreases and the showers decrease in intensity, and the cumulus clouds begin to flatten out into bulging layers resembling stratus, strato-cumulus, or nimbo-stratus. This influence is first felt in the lower layers and gradually spreads upward.

When a continental mass invades an ocean in summer, it develops into a stable state. When a continental mass invades an ocean in winter, its instability increases and the showers increase rapidly in intensity and in frequency.

The shower activity in cold air masses counteracts the instability that is caused by the heating from below. This condition is due to the liberation of the latent heat of vaporization above the condensation level. Therefore, old cold masses gradually develop toward a stable state of equilibrium. It happens, therefore, quite frequently that a cold mass that has traveled from the United States across the Atlantic Ocean arrives in Western Europe in a fairly stable state. The maximum instability occurs in air masses that have moved rapidly directly from the polar regions.

The Properties of a Warm Mass.—By far the most important sources of warm air masses are the subtropical anticyclones in oceanic regions. In summer, warm masses may also develop over southern continents, notably in anticyclonic situations.

The air in the maritime subtropical anticyclones is warm and fairly stable, and the moisture content is high, particularly in the lower layers. When such a mass travels toward colder regions, it arrives there with a temperature higher than that of the surface over which it travels. The air becomes cooled from below, and the lower layer of air becomes increasingly stable. The pronounced stability that develops in this way hinders turbulence and completely prevents convectional currents. The result of this process is that the cooling is mainly limited to the lower layer of the air mass. The cooling from below results in a cooled surface layer of air, the air above the inversion being mainly unaffected, except for the slow cooling caused by outgoing radiation.

By continued cooling from below, the air along the surface of the earth may be cooled below its dew point, in which case fog forms. If the wind velocity is high, there may be sufficient mechanical turbulence to stir the air under the inversion (see Fig. 48); stratus, and not fog, would then develop, as has been explained in a previous section.

A typical maritime warm mass is recognized by the presence of several or all of the following characteristics:

1. Stable lapse rate or inversions in the lower layers.
2. Slight turbulence. Steady wind.
3. Poor visibility.
4. High relative humidity.
5. Stratus, mist or fog, and haze.
6. Drizzle.

Air from the doldrums or from the western part of subtropical anticyclones and warm air that develops over continents in summer are usually conditionally unstable. When such air moves toward colder regions, it will be stable in the lower layer but will remain conditionally unstable at greater altitudes. It will, therefore, have properties characteristic of a warm mass near the earth's surface, but cold-air-mass properties above. This is typical of the tropical air that invades the United States in winter.

When a warm air mass invades a warm continent in summer, instability develops rapidly; the air changes from a warm mass to a cold mass. In winter, on the other hand, the stability is increased and deep layers of fog (advection fog) may cover large areas.

Examples of Cold and Warm Air Masses.—Figure 82 shows a simplified weather map with a cold mass of polar air streaming toward the southeast and meeting with a warm mass of tropical maritime air from the southwest. The "fronts" indicate the lines of separation between the two air masses. It will be seen that cumulus clouds and showers are typical of the cold mass,

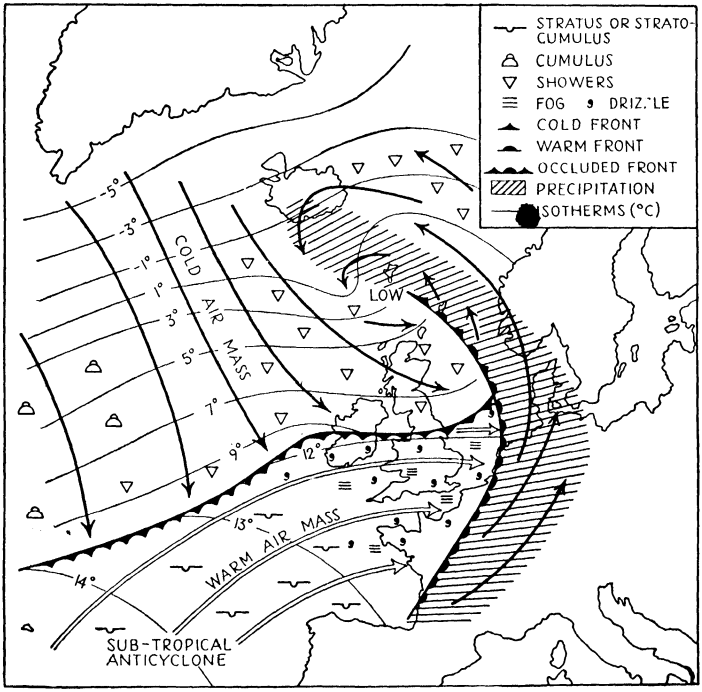


FIG. 82.—Example of warm and cold air masses.

whereas stratiform clouds characterize the warm mass. The northeastern part of the warm mass has traveled the longest distance over a colder surface. As a result, the air has cooled, and fog (advection fog) has formed. In places, drizzle falls from the fog and low stratus.

Figure 83 shows some typical examples of the clouds and weather phenomena that develop within stable and unstable air masses under various conditions.

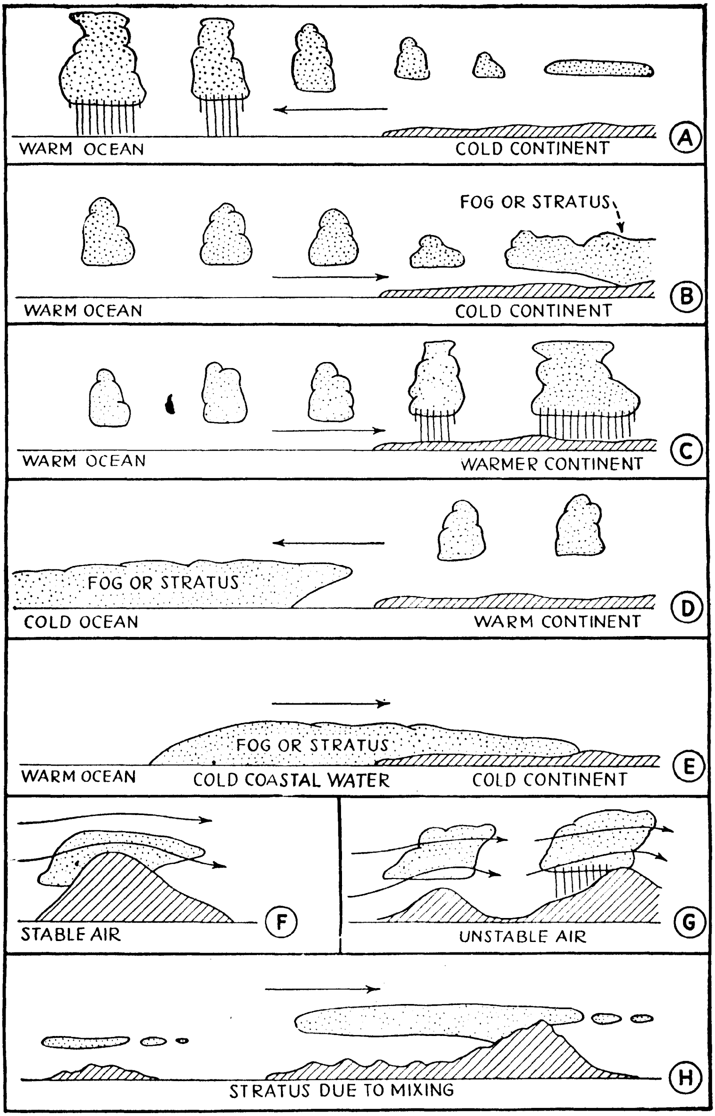


FIG. 83.—Diagrams showing types of clouds in air masses that are heated or cooled from below. Arrows indicate direction of air motion.

CHAPTER IX

FRONTS

So far we have discussed in outline the processes and weather phenomena that develop within the air masses. We shall now proceed to discuss the phenomena that occur on the border between adjacent air masses. The importance of this discussion is readily appreciated when we consider the fact that the colder air mass is denser than the warmer, with the result that a cold and warm air mass in juxtaposition represent a source of potential energy as well as heat energy which may be utilized to create kinetic energy. Furthermore, since the warm air tends to ascend above the cold air, condensation will occur, with the result that the latent heat of vaporization is liberated. From these sources of energy, kinetic energy (wind) is created.

Frontogenesis.—Certain types of motion (see Fig. 79) have a tendency to bring air masses from different and distant source regions into juxtaposition. The air that travels from a colder to a warmer region will be heated, and the air that travels in the opposite direction will be cooled; but if the motion is sufficiently fast, a temperature contrast may develop along the line where the two opposing air currents meet. The term *frontal surface* has been introduced into meteorological literature to denote the surface of separation between two adjacent air masses of different temperature. The line of intersection between the frontal surface and the earth surface is called a *front*, and the process that leads to the formation of a front is called *frontogenesis*. Under certain circumstances, a front may dissolve; this process is called *frontolysis*.

A mathematical analysis of fluid motion shows that the observed motion may be regarded as composed of four elementary components, *viz.*, (1) translation, (2) rotation, (3) divergence, and (4) deformation. These components are shown in Fig. 84. By superimposing these four types and giving each appropriate values, any type of observed motion may result.

What we are here concerned in discussing is the type of motion that can bring distant air masses together so that a discontinuity results.

1. *Translation.*—The type of motion shown in Fig. 84*A* represents the case when the speed and the direction of the wind are constant horizontally. The individual units of air will then maintain their mutual distances, and the isotherms do not tend to approach one another. It follows, then, that this type of motion

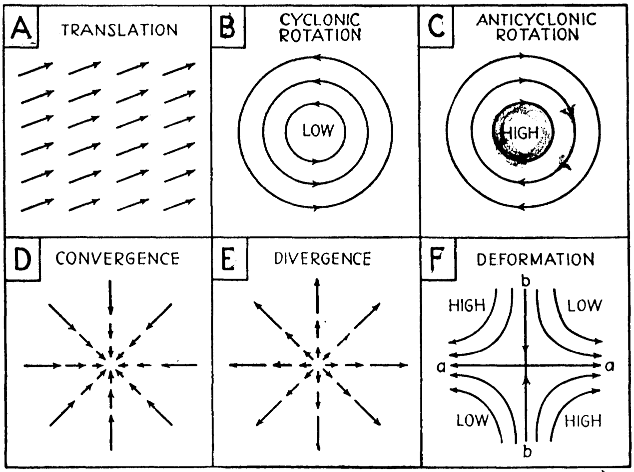


Fig. 84.—Types of motion. Only deformation and convergence contribute to frontogenesis.

cannot create a discontinuity in the temperature distribution. Consequently, the translatory motion is neither frontogenetical nor frontolytical.

2. *Rotation.*—The type of motion shown in Fig. 84*B* and *C* represents cyclonic and anticyclonic rotation, respectively. The streamlines are here concentric circles, and the speed is constant along each circle. The individual units of air will therefore maintain their initial distances, and no discontinuity in temperature will result. Consequently, the rotational motion is neither frontogenetical nor frontolytical.

3. *Convergence and Divergence.*—These types of motion are shown in Fig. 84*D* and *E*. It will be seen that, in the case of convergence, the distance between the individual units of air

will tend to decrease and that, as the moving air (to a certain extent) carries its temperature with it, temperature contrasts may develop. The opposite is true of divergence. However, the temperature contrasts that could be created in this way would not be arranged along a line as is required for the creation of a front

It should be noted that motions of the types shown in Fig. 84*D* and *E* cannot exist alone on a rotating earth, for there are no isobars that correspond to this motion. In small amounts, convergence and divergence may be superimposed on other types of motion, causing the wind to blow slightly across the isobars. Since convergence and divergence are present only in small

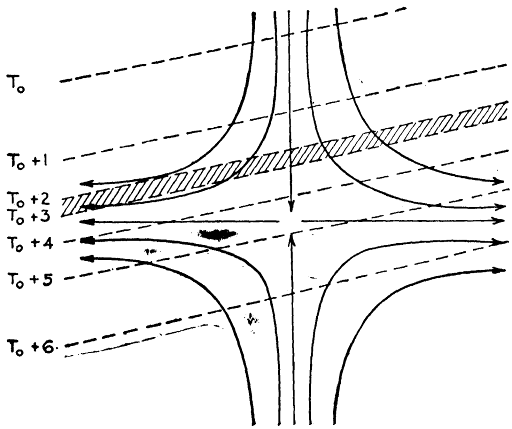


FIG. 85.—Illustrating the movement of the isotherms in the vicinity of a col. This case gives frontogenesis.

amounts and since they have no tendency to accumulate the isotherms along a line, they are relatively unimportant as far as frontogenesis and frontolysis are concerned.

4. *Deformation.*—The type of motion shown in Fig. 84*F* corresponds to what would be observed in the vicinity of a col (see page 108). Here two opposing currents meet and spread out laterally along the line *aa* which is called the *axis of dilatation.* The other axis *bb* is called the *axis of contraction.* To simplify, we shall call *aa* the *axis of outflow* and *bb* the *axis of inflow.*

To study in greater detail the behavior of the isotherms in a field of deformation, we consider Fig. 85. It will be seen that the isotherms will tend to approach the axis of outflow, along which a temperature discontinuity will develop if the motion persists. The isotherms will also tend to rotate and become parallel to the axis of outflow. Figure 85 represents a typical case of frontogenesis.

Let us next consider Fig. 86 in which the angle between the isotherms and the axis of outflow is larger than 45°. The point A will move upward to the center of the col, and the point B will move to the right. Thus the isotherms will tend to rotate and become parallel to the axis of outflow. But as long as the angle Ψ in Fig. 86 is greater than 45°, the distance between neighboring

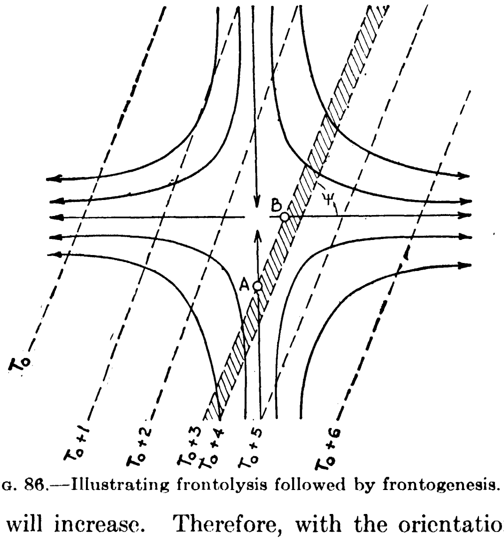


Fig. 86.—Illustrating frontolysis followed by frontogenesis.

isotherms will increase. Therefore, with the orientation of the isotherms shown in Fig. 86, the isotherms will not approach one another until the angle Ψ has become less than 45°. Figure 86 represents, therefore, a case of frontolysis that gradually may develop into frontogenesis.

It follows from the above discussion that the air motion in the vicinity of a col is frontogenetical only when the angle between the axis of outflow and the isotherms is less than 45°. The smaller this angle, the more rapidly are the isotherms concentrated in the vicinity of the axis of outflow.

It should be noted that the intensity of the frontogenetical process is directly proportional to the temperature gradient and the speed with which the air moves toward the axis of outflow. Active frontogenesis, therefore, occurs only when the temperature gradient and the transport of air are sufficiently strong.

5. *Types of col.*—The types of col discussed so far represent *pure deformation;* the angle between the axis of outflow and the axis of inflow is then 90° (see Fig. 87*A*). If a slight anticyclonic rotation is superimposed, we obtain a col of the type shown in Fig. 87*B*; and if a slight cyclonic rotation is superimposed, we obtain the type shown in Fig. 87*C*. Experience shows that the "cyclonic" type of col (*i.e.*, Fig. 87*C*) gives more effective fronto-

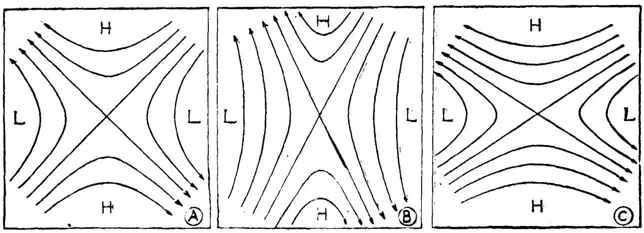


FIG. 87.—Three types of cols: *A*, without rotation; *B*, anticyclonic rotation; *C*, cyclonic rotation.

genesis than do the other types. Least effective is the "anticyclonic" type (*i.e.*, Fig. 87*B*).

The Principal Frontal Zones.—The mean position of the principal frontal zones follows directly from mean-wind charts (Figs. 73 and 75) and the mean distribution of air masses (Figs. 80 and 81). Maps showing the principal frontal zones in winter and summer are seen in Figs. 88 and 89. In winter the Atlantic polar front separates the polar continental air of North America from the tropical maritime air of the North Atlantic. This frontal zone oscillates within wide limits; although it is generally found off the Atlantic coast, it may move northward to southeastern Canada, and it is frequently prolonged eastward to Western Europe. Most of the winter precipitation in the eastern part of the United States is associated with this frontal zone.

In summer (see Fig. 89), this frontal zone moves northward with the prevailing winds (monsoon) and is found in most cases along the southern border of Canada. In summer the temperature contrasts are small and the front is not nearly so well defined as during the winter season.

In winter a frontal zone exists along the Rocky Mountains and the coast of Alaska, separating polar maritime air from polar continental air.

The Atlantic arctic front separates the maritime polar air of the North Atlantic from the arctic air. This frontal zone, too, decreases in intensity in summer.

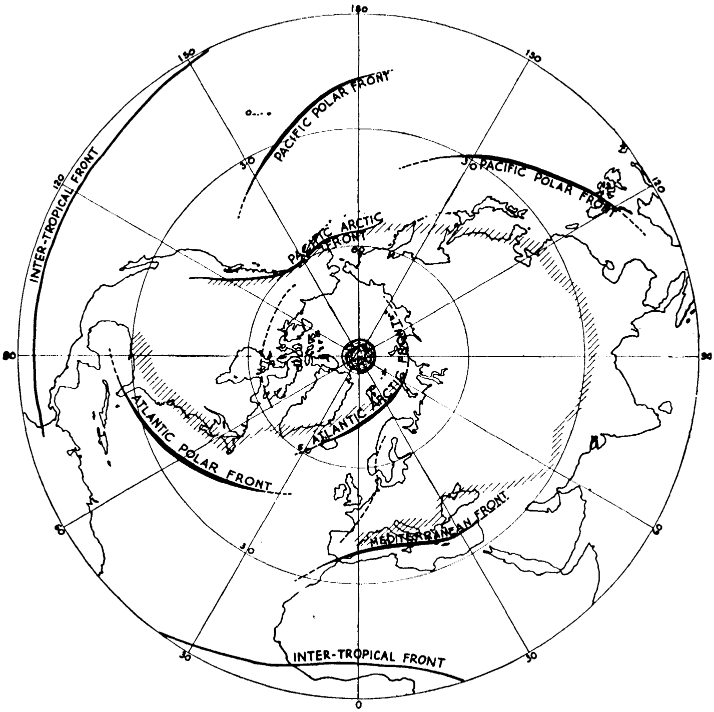


FIG. 88.—The principal frontal zones in winter.

The conditions on the Pacific Ocean are similar to those over the Atlantic Ocean. A pronounced frontal zone is found in winter off the Asiatic coast, separating the cold air of continental origin from the maritime tropical air of the North Pacific. Quite frequently, the subtropical anticyclone over the Pacific breaks up into two separate cells, and a frontal zone forms through the col between them. In summer, the monsoon system in east Asia brings the frontal zone northward to the east coast of Siberia.

During the winter season, a frontal zone is usually found over the Mediterranean, separating the cold continental air over Europe from the warm tropical air over north Africa. This zone vanishes completely in summer and accounts for the fact that the

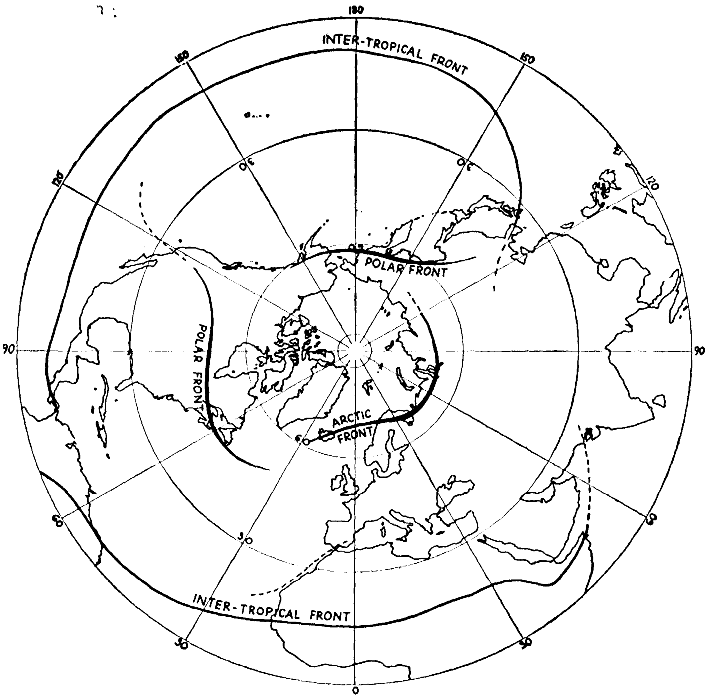


FIG. 89.—The principal frontal zones in summer.

summer season in the Mediterranean is dry and the winter season is rainy.

The pronounced convergence of air into the doldrums tends to create a frontal zone, as shown in Figs. 88 and 89. Since the temperature contrasts are small in the equatorial region, this frontal zone is weak. However, the general convergence of the flow gives rise to ascending air currents which account for the heavy precipitation that is characteristic of the equatorial belt.

It will be seen from Figs. 88 and 89 that the equatorial frontal zone moves northward in summer and southward in winter. In the Indian region the summer monsoon brings the frontal zone into the interior of Asia, and the winter monsoon sweeps it to the south of the equator.

The significance of these frontal zones lies in the fact that, on account of the temperature contrasts associated with them, they represent the main source of energy for the creation of storms.

The vertical extent of the frontal zones is shown diagrammatically in Fig. 90. The colder air forms a wedge under the warmer air which tends to ascend along the frontal surface. The

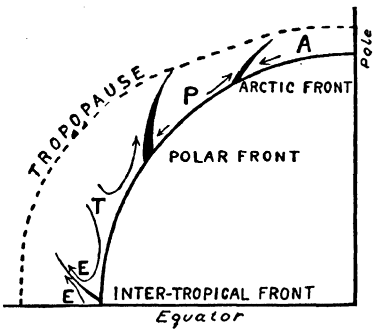


FIG. 90.—Diagrammatical cross section showing the principal frontal zones and air masses in the Northern Hemisphere.

vertical extent of the frontal zones may vary within wide limits. On the whole, we may say that air masses which are formed through heating from below are deep masses, whereas those which form through cooling from below are shallow. The fronts that separate tropical air masses from colder air have, therefore, the greatest vertical extent.

Inclination of Frontal Surfaces.—Since a frontal surface is a discontinuity in temperature, it will also be a discontinuity in density, the colder air being denser than the warmer air. The surface of separation between a dense and a less dense fluid will tend to become horizontal. If the earth did not rotate, the surface of separation would be exactly horizontal in the equilibrium state. However, on account of the earth's rotation, a frontal surface will be in an equilibrium position when it is slightly inclined to the horizontal. The inclination depends also upon the density discontinuity and the wind distribution; but, under normal conditions, the inclination of the frontal surfaces in middle latitudes is approximately $1/100$. Although the inclination is slight, it is nevertheless important. In all diagrams showing cross sections through frontal surfaces, the vertical scale is exaggerated relative to the horizontal scale to bring out the slope clearly.

Fronts in Relation to Temperature.—If the frontal surface were a perfect discontinuity, a temperature-height curve would be as

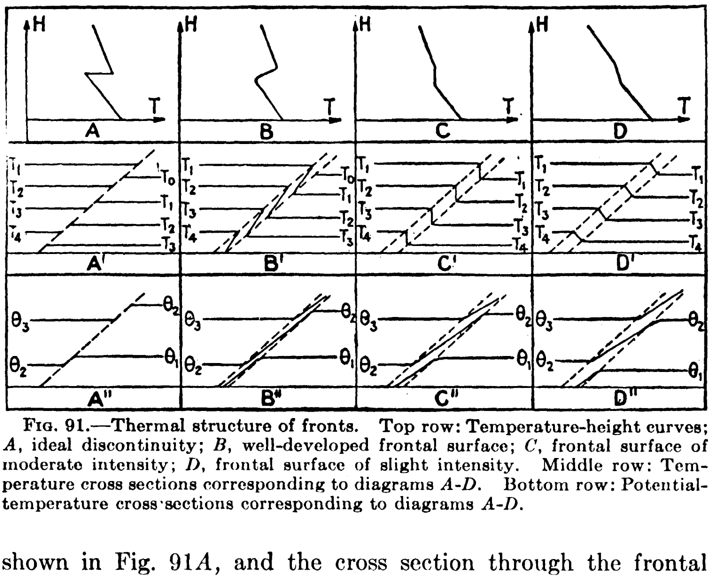


FIG. 91.—Thermal structure of fronts. Top row: Temperature-height curves; *A*, ideal discontinuity; *B*, well-developed frontal surface; *C*, frontal surface of moderate intensity; *D*, frontal surface of slight intensity. Middle row: Temperature cross sections corresponding to diagrams *A-D*. Bottom row: Potential-temperature cross sections corresponding to diagrams *A-D*.

shown in Fig. 91*A*, and the cross section through the frontal surface would be as shown in Fig. 91*A'*. However, perfect discontinuities are never found in the atmosphere; instead, we find a relatively narrow layer of transition between the cold and the warm air masses, the layer of transition being mainly due to mixing. A well-developed frontal surface will then have a thermal structure, as shown in Fig. 52. This is indicated diagrammatically in Fig. 91*B* and *B'*. If the front weakens, the temperature contrasts would be less pronounced (Fig. 91*C* and *C'*), and a very weak front would have a temperature distribution along the vertical as shown in Fig. 91*D* and *D'*.

The lower portion of Fig. 91 shows the distribution of potential temperature along a frontal surface. The potential-temperature lines have a tendency to follow the frontal surface.

The thickness of the layer of transition varies within wide limits. At a well-developed front, it may be only 500 ft.; at diffuse fronts, it may be several thousand feet in thickness. Since the inclination of the frontal surface is about $\frac{1}{100}$, it follows that the width of the frontal zone in the horizontal direction will be about one hundred times larger than the vertical thickness.

Fronts in Relation to Pressure.—We consider the vertical cross section through a frontal surface as shown in Fig. 92 and recall that the atmospheric pressure is equal to the weight of the air

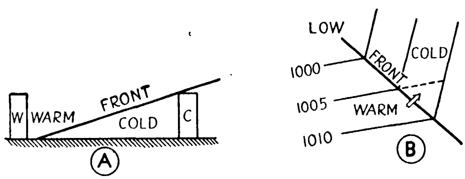


FIG. 92.—Illustrating the refraction of isobars at the frontal surface.

column. Since the cold air is denser than the warm air, it follows that the air column C (in the cold air) must contribute more to the atmospheric pressure at the surface than does the warm air column W of equal height.

Let us next consider Fig. 92B which represents the atmospheric pressure at the ground. We proceed along an isobar (say, 1005 mb.) toward the front. As we enter the cold air, the pressure must increase on account of the excess weight of the colder air, and if we proceed far enough we shall meet with the 1010-mb. isobar. This shows that *the isobars at a front must be refracted in such a manner that the kink in the isobar points from low to high pressure.* This is one of the fundamental frontal characteristics and applies to all types of front. Figure 93 shows the principal types of isobar associated with fronts.

Suppose now that the front moves in the direction of the arrows shown in Fig. 93A. It follows then that the pressure would fall (while the front approaches) and then rise. As the front passes a station, there would be a kink pointing downward in the barogram at the station. Depending on the pressure distribution around the front, the barogram may take various shapes, but there will always be a kink in the barogram with the kink pointing downward; the sharper the front and the greater its speed, the more pronounced the kink.

Fronts in Relation to Wind.—Since the winds near the earth's surface blow mainly along the isobars with a slight drift toward lower pressure, it follows that the wind direction in the vicinity of a front must conform with the refraction in the isobars. The arrows in Fig. 93 indicate the winds that correspond to the pressure distribution. It will be seen that a front is a *wind-shift line* and that the wind shifts in a *cyclonic sense.* Since the front moves in the direction of the wind component normal to the front

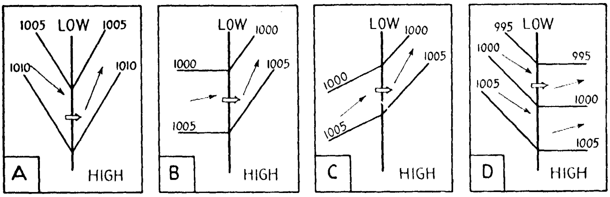


FIG. 93.—Types of isobars associated with fronts.

(*i.e.*, the double-shaft arrows), it is readily seen that the following rule holds for all types of front: *If you stand with your back against the wind in advance of the front, the wind will shift to your left-hand side as the front passes.*

The speed of the wind depends on the pressure gradient. Thus, in Fig. 93*A* the speed of the wind would be about the same in both air masses; in Fig. 93*B* a relatively strong wind would be followed by a weaker wind; and in Fig. 93*D*, a weak wind would be followed by a strong wind.

We next consider Fig. 94. Suppose that we ascend vertically from the cold toward the warm air. In ascending along the vertical *AA*, we first experience the wind in advance of the front and then the winds in the rear of the front; hence, the winds will shift as indicated in the lower right portion of the diagram. However, if we ascend along the line *CC*, we first experience the wind in the rear of the front and then the wind in advance of the front, and the wind direction will shift as shown in the lower left portion of the diagram.

Classification of Fronts.—The frontal characteristics described so far are such as apply to all fronts whether colder air replaces warmer air, or vice versa, or whether the front remains stationary.

These characteristics may, therefore, be called *general front characteristics.* Depending on the movement of the front and the stability conditions of the air masses, a number of additional characteristics may be described. It is, therefore, appropriate to classify the fronts in order to specify the special characteristics that apply to the individual types.

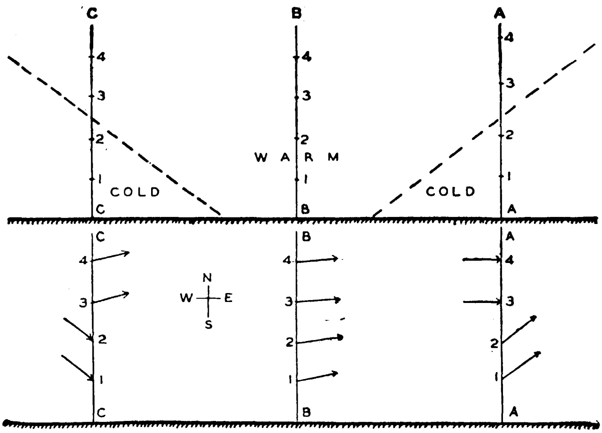


FIG. 94.—Vertical distribution of wind in the vicinity of frontal surfaces.

1. *A cold front* is a front along which colder air replaces warmer air.
2. *A warm front* is a front along which warmer air replaces colder air.
3. *A stationary front* is a front along which one air mass does not replace the other.
4. *An occluded front* is a front resulting when a cold front overtakes a warm front.

Although the difference in temperature (or density) is the primary factor, the most important characteristic is the cloud systems and the weather phenomena associated with the types of front listed above.

Fronts and Clouds.—The warmer air, being less dense than the colder air, will tend to ascend along the frontal surface and cool adiabatically. If the air is not excessively dry, this will lead to the condensation of the water vapor, and a cloud system develops above the frontal surface. Clouds of secondary importance may form also under the frontal surface, but only those forming above and having an appreciable ascending velocity will contribute materially to the precipitation.

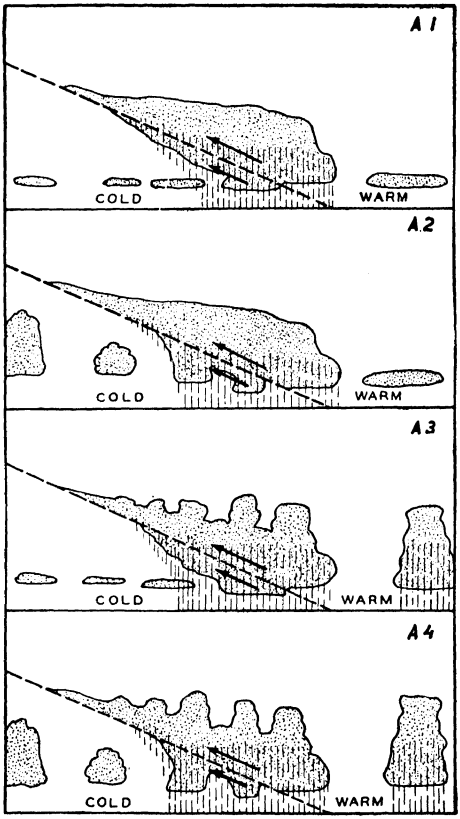


FIG. 95.—Cross sections through warm-front surfaces. *A*1, both masses are stable; *A*2, the warm air is stable and the cold air unstable; *A*3, the warm air is unstable and the cold air stable; *A*4, both masses are unstable.

1. *Warm Fronts.*—We shall first consider the warm fronts as shown in the vertical cross sections in Fig. 95. Here, the fronts move from the warm toward the cold side, or from the right to the left.

The four diagrams in Fig. 95 represent different types of warm front, but in all cases the following is true: The warm air ascending above the cold wedge will produce a vast cloud system, the width of which may vary from, say, 100 to 300 miles or more, and the length of which may be 1000 miles or more. The height of the cloud system may vary within wide limits from about 6000 to about 20,000 ft. Thus, the warm fronts produce extensive cloud systems of great vertical extent. Usually, the clouds reach well into the layers where subfreezing temperatures occur. The upper portion of the cloud system will therefore normally contain ice particles, and this causes precipitation to be released, as was explained on page 46. A more detailed picture of the structure of such a front is shown in Fig. 33.

The normal sequence of clouds, from the upper to the lower portion of a warm front, is as follows: *cirrus, cirro-stratus, alto-stratus, nimbo-stratus* (see Figs. 18, 19 and 21).

Thus, when a warm front approaches, an observer on the earth's surface will see the clouds arriving in the above sequence, the various types merging gradually into one another. Low clouds of the stratus or cumulus type may be present under the front so that the upper clouds are partly obscured.

In detail, the warm-front cloud system varies with the stability conditions of the air, as shown in the four diagrams in Fig. 95.

If both air masses are stable, we obtain a "smooth" cloud system as shown in diagram *A*1. Low stratus may or may not be present in the warm air, and stratus may form also in the cold air; but the typical warm-front cloud system is essentially as described above.

If the warm air is stable and the cold air unstable, we obtain a cross section as shown in diagram *A*2.

If the warm air is unstable or convectively unstable, convective clouds will form and grow up from the warm-front cloud system as shown in the diagrams *A*3 and *A*4.

When the warm air is stable (diagrams *A*1 and *A*2), the precipitation falls evenly without sudden changes. However, when the warm ascending air is unstable, showery or squally precipitation is superimposed on the more even frontal precipitation. Under such conditions, thunderstorms may occur as a result of convection in the warm air above the frontal surface (warm-front thunderstorms).

2. *Stationary Fronts.*—What has been said above of warm fronts applies also to stationary fronts. Although the stationary

fronts do not move, the warm air will ascend and the cloud systems will assume the forms shown in Fig. 95.

3. *Cold Fronts.*—Figure 96 shows cross sections through cold fronts under various conditions. Here the fronts move from the

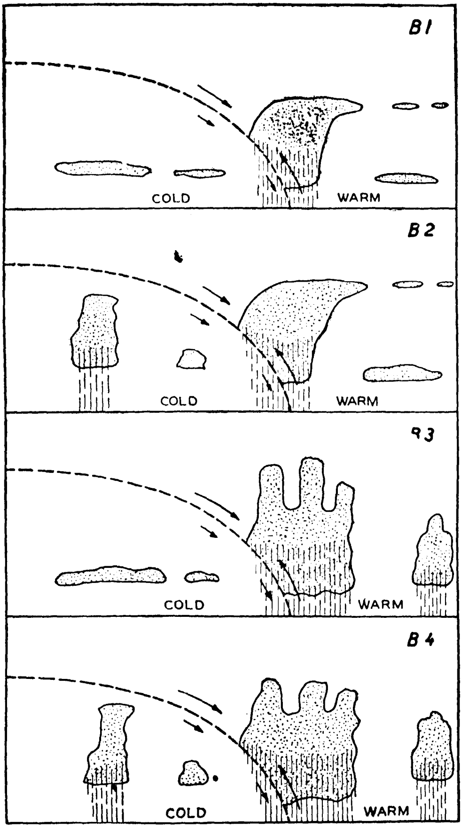


FIG. 96.—Cross sections through cold fronts. *B*1, both masses are stable; *B*2, the warm air is stable and the cold air is unstable; *B*3, the warm air is unstable and the cold air stable; *B*4, both masses are unstable.

cold to the warm side, *i.e.*, from the left to the right. It will be noticed that the cold-front surface is steeper than the warm-front surface. This is mainly due to the fact that the wind velocity increases along the vertical (see Fig. 66), with the result that the front will tend to move faster in mid-air than at the ground.

In detail the cloud system varies with the stability conditions of the air masses, but the following features hold in all cases: The warm air above the cold-front surface, moving faster than the cold air, will descend along the upper portion of the cold front. The descending air heats up adiabatically and becomes relatively dry. Thus, usually, there is no cloud system above the upper portion of the cold front. The warm air at low levels moves slower than the front and will therefore ascend. This causes a cloud system above the cold front with a tendency to tilt in a forward direction.

The cloud systems connected with typical cold fronts vary in appearance with the stability of the air, as shown in Fig. 96. If the warm ascending air is stable, a "smooth" cloud system (Fig.

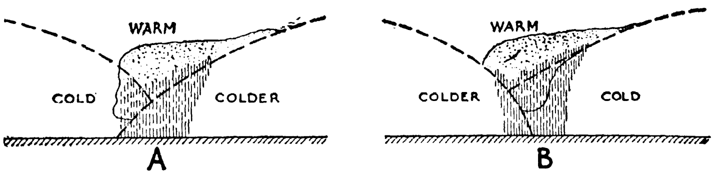


FIG. 97.—Cross sections through occlusions: *A*, warm-front type occlusion; *B*, cold-front type occlusion.

96*B*1 and *B*2) results, with a more or less even fall of precipitation. However, if the warm air is unstable or convectively unstable, squally precipitation is superimposed. Frequently thunderstorms develop in connection with cold fronts, and in cases of extreme instability the typical cold-front cloud is of the cumulonimbus arcus type (Fig. 28); the cold front is then called a *squall line*.

The diagrams shown in Fig. 96 are typical of cold fronts that move with moderate or great speed. If the cold front moves slowly, there is less descending current above the upper portion of the frontal surface, and the cloud system extends far in the rear of the front at the ground. When the cold front moves very slowly and approaches the quasi-stationary state, the cross section through the cloud system will be similar to that shown in Fig. 95.

4. *Occluded Fronts.*—An occluded front results when a cold front overtakes a warm front. The cloud system that accompanies an occluded front will therefore be a combination of the two, as shown in Fig. 97. In these diagrams the frontal system

moves from the left toward the right. The warm air that originally was between the warm and the cold front has been squeezed upward. If the air in advance of the occluded front is colder than that in the rear, we obtain the conditions represented in Fig. 97*A*. This is called a *warm-front type occlusion.* If the rear air is the coldest mass, we have Fig. 97*B*, which represents a *cold-front type occlusion.* The sequence of clouds during the approach of an occluded front is essentially the same as that associated with warm fronts.

Fronts and Wind Structure.—A passage of a well-marked front is often accompanied by gusts and wind squalls, which, in violent cases, may cause difficulties for aircraft operations. These gusts and squalls depend mainly on three factors: (1) the speed of the wind, (2) the stability conditions in the air masses, and (3) distribution of vertical velocity at the front.

The speed of the wind is proportional to the pressure gradient. Thus, a frontal squall is to be expected at the passage of a front when the pressure gradient in the rear of the front is larger than that in advance of the front. Such squalls are always associated with a wind shift of cyclonic character (*i.e.*, veering). Conversely, a lull would occur at the passage of the front if the pressure gradient in the rear is weaker than in advance of the front.

The intensity of turbulence depends on the speed of the wind and the stability conditions of the air masses. Other conditions being equal, the turbulent gusts are most intense in the air mass that is least stable. Thus, strong gusts are likely to occur along fronts where rapidly moving unstable air replaces slowly moving stable air. Such conditions occur most frequently along cold fronts with northerly or northwesterly currents in the rear.

The intensity of the frontal squalls that occur near the ground depends also on the distribution of vertical velocity along the frontal surface. The reason for this is best understood by comparing Fig. 95 with Fig. 96. If the air ascends along the frontal surface in both masses (Fig. 95), there must be a convergent flow toward the front. The air that arrives at the front at the ground will then be exposed to the frictional drag along the earth's surface for an extended interval of time; as a result, its speed will be much below the geostrophic value. However, when the air in the rear of the front descends (Fig. 96), it will come from higher levels where the speed is close to the geostrophic wind. From the above, it follows that frontal wind squalls of considerable intensity may occur during the passage of cold fronts, notably when the cold air is unstable and the pressure gradient in the cold air is greater than that in the warm air.

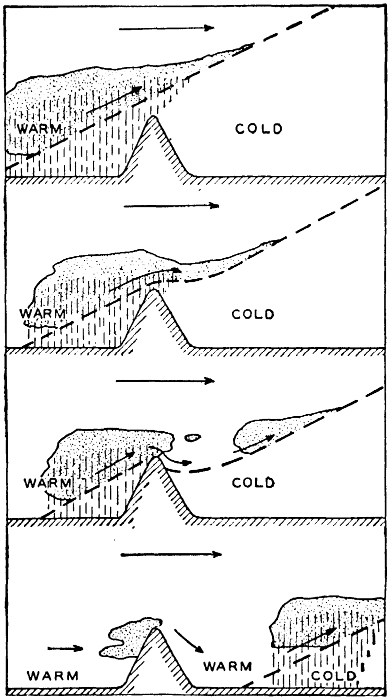


FIG. 98.—Successive cross sections through a warm-front surface passing a mountain range.

Influence of Mountain Ranges.—The frontal cloud systems are disturbed to a considerable degree when the fronts pass a mountain range. Figure 98 shows cross sections through a warm-front surface that passes a mountain range. The descending current that develops on the lee side tends to dissolve the cloud system above the lee slope of the range, and the movement of the frontal surface is retarded on the windward side. As a result, most of the precipitation falls on the windward side of the range.

Figure 99 shows the passage of a cold front across a mountain range. As in the previous case, most of the precipitation falls on the windward side.

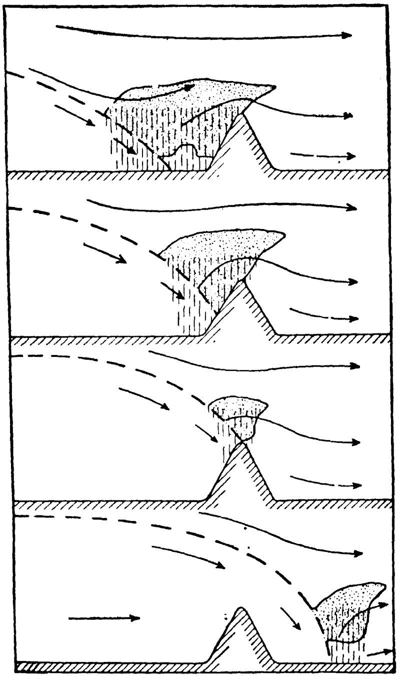


FIG. 99.—Successive cross sections through a cold-front surface passing a mountain range.

The influence of mountain ranges on the distribution of precipitation is evident from Fig. 139, which shows the annual mean amount of precipitation. Note the large amounts of precipitation on the windward slope in the prevailing westerlies and in the monsoon region.

CHAPTER X

CYCLONES AND ANTICYCLONES

As mentioned on page 108, a cyclone may be defined as a region of low pressure surrounded by closed isobars and an anticyclone as a region of high pressure surrounded by closed isobars. Prior to 1918, a cyclone was thought of merely as a region of relatively low pressure usually associated with bad weather. About 1918, J. Bjerknes examined a large number of cyclones by means of observations from a dense network of stations. He then found that the distribution of wind, temperature, humidity, etc., was not continuous but discontinuous in most cyclones. By studying a large number of cyclones, he was able to make a *cyclone model* which shows the essential features of a cyclone. This model may be taken as a blueprint, showing the essential features of any cyclone as it appears on a weather map.

The Cyclone Model.—Figure 100 shows the cyclone model. The middle portion of the diagram represents the cyclone as it appears on a weather map at sea level. The lower portion of the diagram represents a vertical cross section running from west to east to the south of the center, and the upper portion represents a similar cross section to the north of the center. We shall first discuss the middle portion of Fig. 100.

Bjerknes found that a cyclone normally consists of two air masses separated from one another by a front. A tongue of warm air, usually from the south, extends towards the center of the cyclone where the pressure is lowest. This tongue of warm air is called the *warm sector*.

The warm sector, Bjerknes found, is usually surrounded by cold air of polar origin, the warm air being usually of tropical origin. Thus, the cyclone is built up of two air masses of widely different temperature and life history. The front that runs through the cyclone was called the *polar front*, since it most frequently represents the southward border of the polar air masses. Along the right-hand portion of the front, warmer air replaces colder air;

this part of the front is called a *warm front.* To the left, colder air replaces warmer air; this part of the front is called a *cold front.*

Furthermore, it was found that the principal cloud system and area of precipitation are arranged as a band along the fronts on their colder side and that the center of the cyclone usually moves as indicated by the broken arrow.

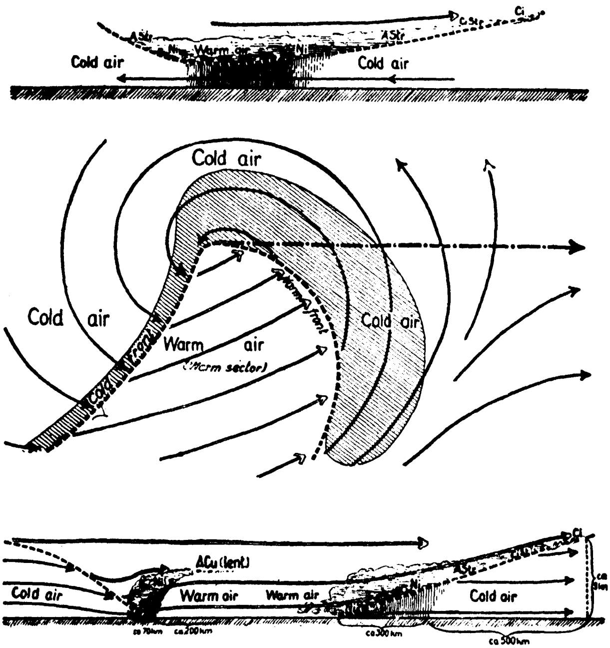


FIG. 100.—The cyclone model. (*After J. Bjerknes and H. Solberg.*)

We shall next consider the lower portion of Fig. 100, which represents a cross section through the cyclone south of its center. It will then be seen that the cold air in advance of the warm front forms a wedge under the warmer air above. The warm air, being lighter than the cold air, ascends up the slope; as it does so, it comes under lower pressure and cools adiabatically, and this results in the condensation of the water vapor. Thus the cloud system that is indicated in advance of the warm front in the middle portion of the diagram actually rests on the wedge of cold air. At the uppermost portion of the warm-front surface the clouds are high and thin (cirrus and cirro-stratus). As we go down the slope, the clouds become denser (alto-stratus) and merge gradually into the rain cloud (nimbo-stratus).

In the rear of the cold front, the wedge of cold air pushes under the warmer air, and a cloud system results through lifting and cooling of the warm air. This cloud system is usually not so wide as is the warm-front system. However, the cold-front cloud systems may vary considerably, as was explained in the preceding chapter.

The upper portion of Fig. 100 shows a cross section north of the center. Here the warm air does not come down to the earth's surface but is present aloft. If we ascended from the ground north of the cyclone center, we should have easterly winds in the cold air and, as we go through the frontal surface, we should meet with winds from a westerly direction. South of the center, the wind variation with height would be as shown in Fig. 94.

Bjerknes's cyclone model indicates not only the structure, or the anatomy, of a cyclone but also the sources of energy. Thus the juxtaposition of a cold and a warm mass side by side represents potential as well as heat energy from which kinetic energy may be created. Furthermore, since condensation occurs in the ascending air, the latent heat of vaporization is liberated and adds to the available energy.

For some time, it was thought that all cyclones conform with the model shown in Fig. 100. However, later investigations soon showed that the cyclone model represents only a single, but important, stage in the life history of a cyclone.

Stable and Unstable Waves.—Before we proceed to discuss the development of a cyclone, we shall consider some elementary aspects of wave motion on the boundary between a dense and a less dense fluid. The swell on an ocean represents a simple type of wave motion. Here, the water is the denser, and the air the less dense medium. The waves are characterized by a regular profile, their amplitudes remain almost constant, and the surface bulges up and down. The particles at the wave crest have a maximum of potential energy, and those in the wave trough have a minimum of potential energy. The wave-generating force is gravity, and potential energy is converted into kinetic energy, and vice versa. Waves of this type are called *stable waves* because their amplitude does not change with time.

We shall next consider wind waves on a water surface. If the wind velocity is slight, the waves are essentially stable. However, if the wind increases and surpasses a certain value, the wave crests grow and become tall and slim, and eventually the wave crest breaks. A wave of this kind is *unstable*, and the force that creates instability is the *wind shear* at the water surface.

Now, a frontal surface in the atmosphere separates a denser mass from a less dense mass. The difference in density between two adjacent air masses is, of course, small, but it is nevertheless significant. Furthermore, a frontal surface is not horizontal; it has an inclination of about $\frac{1}{100}$, as was explained in the section on fronts (see Fig. 91). The wave-generating force on a frontal

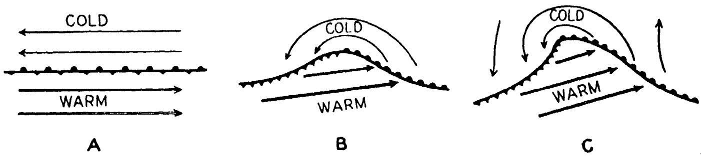


FIG. 101.—Development of a wave cyclone prior to the stage represented by the cyclone model.

surface is partly due to gravity and partly due to the inertia caused by the rotation of the earth. Thus, the gravity force and the inertia force tend to create waves on frontal surfaces; and if the wind shear is sufficiently strong, the waves will be unstable. The inertia force is unimportant in motion on a small scale, but important in motion of the dimensions of a cyclone.

The Development of Cyclones.—We shall now endeavor to explain the development of a cyclone prior to the stage represented by the cyclone model (Fig. 100), and also the development that follows after this stage.

We consider first Fig. 101*A*. Here we have, say, a westerly current of warm air next to, say, an easterly current of cold air. The front is stationary because the air streams parallel to the front. It is important to note that the cold air forms a wedge under the warm air, the frontal surface having an inclination of about $\frac{1}{100}$. As we ascend through the cold air and pass the frontal surface, the wind changes suddenly from an easterly to a westerly direction. Thus, there is a strong wind shear at the frontal surface. In the same way as unstable waves form on an ocean surface when there is sufficient wind shear, waves will form on the frontal surface.

Through wave motion, the frontal surface will bulge up and down, and as it does so the front at the ground will oscillate horizontally as shown in Fig. 101.

Figure 101*B* shows a wave that has just been created. Since the waves are unstable, their amplitude will increase with time and will soon reach the stage shown in Fig. 101*C*, which corresponds to the cyclone model. It will be seen that the cyclone commences as a slight wave on a stationary or almost stationary front; its amplitude increases, and after a period of time of about 12 to 24 hr. the wave has developed to the stage represented by the cyclone model. The wave theory of cyclones shows that the

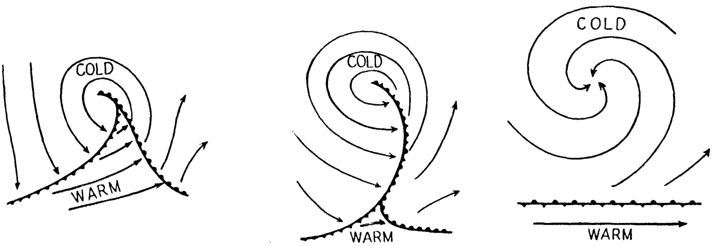


FIG. 102.—Late stages in the development of a cyclone.

waves are unstable only within a certain range of wave length; waves shorter than about 400 miles (600 km.) are stable waves and do not develop. The same is true of waves longer than about 2000 miles (3000 km.). Waves whose lengths lie between about 400 and 2000 miles are unstable when there is sufficient wind shear at the front. Only the unstable waves develop into cyclones; these waves are therefore called *cyclone waves.*

The cyclone wave continues to develop after it has reached the cyclone-model stage. The amplitude continues to increase, the cold front overtakes the warm front, and an *occluded front* results. This development is shown in Fig. 102. As the occlusion process continues, the occluded front dissolves, and the cyclone develops into a large whirl of more or less homogeneous air.

Vertical cross sections through occluded fronts are shown in Fig. 97. If the air in the rear of the cold front is colder than the air in advance of the warm front, a cold-front type occlusion develops. If the air in advance of the warm front is colder than the air in the rear of the cold front, a warm-front type occlusion results, as was explained in the foregoing chapter.

The development from the initial wave disturbance to the cyclone-model stage usually requires only 12 to 24 hr., whereas the development that follows the cyclone-model stage usually lasts for 2 to 3 days or more. Thus, during the major part of its life history, the cyclone is *partly occluded.* It is usually found that the cyclone attains a maximum of intensity (strongest winds) 12 to 24 hr. after the occlusion process has commenced. In the later stages of its development, as the occlusion process proceeds

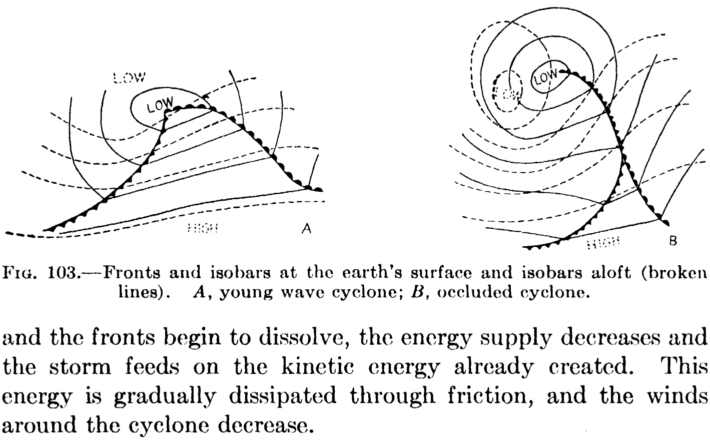


FIG. 103.—Fronts and isobars at the earth's surface and isobars aloft (broken lines). *A*, young wave cyclone; *B*, occluded cyclone.

and the fronts begin to dissolve, the energy supply decreases and the storm feeds on the kinetic energy already created. This energy is gradually dissipated through friction, and the winds around the cyclone decrease.

Observations from the free atmosphere show that the cyclone develops first in the lower atmosphere and gradually builds up to higher levels. Figure 103*A* shows diagrammatically the fronts and the isobars at the earth's surface and the isobars aloft (broken lines) in a young wave cyclone. It will be seen that there is no pressure center aloft, the isobars having a wavelike shape with a trough in the rear of the cold front and a wedge in advance of the warm front. As the wave develops further and occludes (Fig. 103*B*), a pressure center develops aloft, rather in the rear of the center at the ground. As the occlusion process continues, a cyclone center with closed isobars will be found at higher and higher levels.

The disturbance caused by a cyclone affects the tropopause and the lower stratosphere. Figure 104*A* shows a cross section through a young wave cyclone, and Fig. 104*B* a cross section through an occluded cyclone. It will be seen that the cyclone wave induces a wave on the tropopause which is out of phase with the cyclone wave. As the cyclone occludes, the wave on the tropopause becomes more and more pronounced.

It will also be seen from Fig. 104 that, in the initial state, cold (dense) and warm (less dense) air masses are arranged side by side, whereas in the occluded stage the denser mass is under the less dense mass. Thus, during the occlusion process, the potential energy of the system has been decreased. The potential

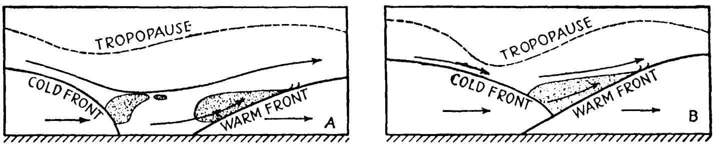


FIG. 104.—A cyclone wave causes a wave on the tropopause whose amplitude increases as the cyclone occludes.

energy thus liberated is one of the principal sources for the creation of kinetic energy.

Tropical Cyclones.—The tropical cyclones are small cyclonic whirls, having nearly circular isobars and very strong winds circulating in a counterclockwise direction in the Northern Hemisphere and in a clockwise direction in the Southern Hemisphere. The tropical cyclones are called *cyclones* in India, *hurricanes* in the West Indies, and *typhoons* in east Asia. They originate in the doldrums over the ocean between 6° and 20°N., or 6° and 20°S., and they travel in the direction of the trade winds along the isobars on the equatorial side of the subtropical anticyclones. At the western end of the subtropical anticyclones, they recurve poleward along more or less parabolic paths, as shown in Fig. 105.

The wind is light and variable in the center (the "eye") of the tropical cyclone around which there is a whirl of hurricane winds and torrential rainfall, often accompanied by thunderstorms. The horizontal diameter of a tropical cyclone varies from a few miles up to several hundred miles. The diameter is smallest when the cyclone is closest to the equator, and it increases as the cyclone recurves poleward. The wind velocity often exceeds

100 m.p.h. The cyclone travels with a moderate speed of about 10 to 20 m.p.h.

The tropical cyclones tend to form when the doldrums are farthest away from the equator. Table VIII shows the percentage frequency of tropical storms in the various regions. It will be seen that no tropical cyclones occur in the South Atlantic. This is due to the fact that the doldrums in this region do not move south of the equator. Because of the pronounced monsoon winds in South Asia, the North Indian Ocean is visited by the doldrums in spring and autumn, in which seasons most tropical cyclones occur. In all other oceanic regions, the highest frequency is observed in late summer and early autumn when the doldrums are farthest away from the equator.

TABLE VIII.—PERCENTAGE FREQUENCY OF TROPICAL CYCLONES

Locality	Jan.	Feb.	Mar.	Apr.	May	June	July	Aug.	Sept.	Oct.	Nov.	Dec.
West Indies	0	0	0	0	1	3	3	26	26	34	4	3
China Sea	0	0	0	2	4	10	19	18	23	13	9	2
Philippines	1	0	1	2	5	9	16	17	19	14	11	5
North Indian Ocean	4	0	3	10	19	10	1	2	3	19	19	5
South Indian Ocean	22	19	18	15	6	1	0	0	0	1	8	10
South Pacific	30	18	28	6	1	0	0	0	1	1	3	12
South Atlantic	0	0	0	0	0	0	0	0	0	0	0	0

There is some evidence that the tropical cyclones in the Atlantic and the Pacific oceans originate along the intertropical front as a disturbance like the cyclones on the polar front. In a well-developed tropical cyclone, no fronts can be traced; for if a front were present originally, it would be destroyed in the intense circulation around the center. On the other hand, there is evidence to show that some of the tropical cyclones in the Indian Ocean do not form along any front, and it is probable that thermal instability plays an important part in the formation of all tropical cyclones.

Cyclone Tracks.—Most cyclones in middle and high latitudes originate as waves along the western portion of the main frontal zones shown in Figs. 88 and 89. In the main, the cyclones travel along these frontal zones with a tendency to curve off toward higher latitudes. Since the frontal zones are not stationary but move north and south with the season and also move irregularly from day to day, the cyclone tracks may vary considerably. Figure 105 gives a highly simplified picture of the principal cyclone tracks.

Most of the cyclones in the North Atlantic originate along the Atlantic coast. This is particularly true during the colder half of the year. Most of these cyclones travel to the northeast and occlude, with the result that most cyclones arriving in Western Europe are occluded cyclones. The majority of the North Atlantic cyclones travel into the Icelandic Low, some travel across the Norwegian Sea into the Barents Sea, and a few may cross the European continent.

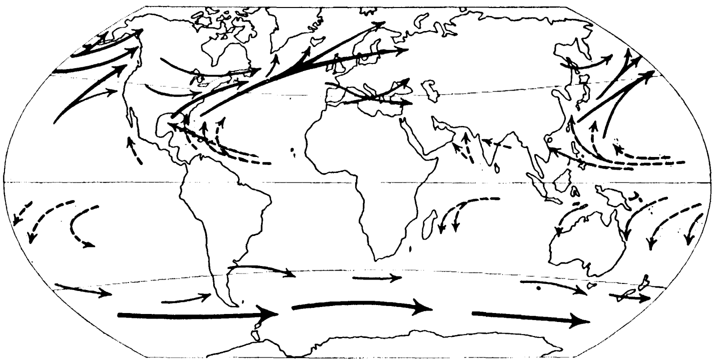


FIG. 105.—The cyclone tracks are highly simplified.

In winter, cyclones form quite frequently along the Mediterranean front. These cyclones usually travel eastward into Asia Minor, and south Russia. A few may cross the mountain ranges and arrive in India. In summer, there is no frontal zone in the Mediterranean region, and, as a result, cyclones are very rare.

Most winter cyclones on the Pacific Ocean originate on the frontal zone off the east coast of Asia, and other cyclones form on the frontal zone that is present frequently over the central part of the North Pacific. The main trend of these cyclones is toward the northeast into the Aleutian Low, where they arrive mostly in the occluded state. Occasionally, these cyclones may cross the Rocky Mountains and arrive in North America. Quite frequently, cyclones form during the winter season along secondary fronts over the North American continent. These, too, travel in an eastward direction. When polar air streams down to the Gulf of Mexico, a strong polar front may form in that region and give rise to cyclones which develop rapidly and travel over the eastern part of North America, giving rise to storms of considerable intensity.

In summer the polar front is displaced farther to the north, and the front is much weaker than in winter. The cyclones are then less pronounced, and the cyclone tracks over the oceans are displaced farther to the north.

In the Southern Hemisphere, most cyclones develop south of 40°S. and travel from the west to the east, as shown in Fig. 105.

The speed with which the cyclones move may vary within wide limits. Over the oceans, a speed of 60 m.p.h. is not infrequent during the winter season, although most cyclones have a speed of about 30 m.p.h. or less.

Tornadoes and Waterspouts.—A tornado is a circular whirl of great intensity and small horizontal extent, in which the wind velocity is usually of superhurricane force. The horizontal diameter of the tornado varies from a few feet up to a mile. The wind velocities sometimes exceed 200 m.p.h. The pressure in the center of the tornado is much lower than in the immediate surroundings; and this, together with the high winds, produces destructive effects. The air in the center is rising rapidly, and the whirl is accompanied by heavy rain or hail and thunder and lightning. The decrease in pressure in the center of the tornado cools the air below its dew point, and, as a result, a funnel-shaped cloud marks the core of the whirl.

The tornadoes are short-lived, usually not lasting more than an hour or two. They usually occur in connection with a strong cold front of the type that produces line squalls. They often form in series and travel in almost parallel paths following the squall line. Tornadoes occur quite frequently in the Mississippi Valley. In Europe, they are rather rare and not so violent as those which occur in the United States.

Waterspouts are tornadoes that form at sea.

Anticyclones.—An anticyclone is an area of high pressure surrounded by closed isobars. The winds blow around the anticyclone in a clockwise direction in the Northern Hemisphere (see Fig. 68) and in a counterclockwise direction in the Southern

Hemisphere. In the center, the winds are light and variable. On the whole, the winds are moderate in anticyclones, except on the outskirts, where the winds occasionally may be strong. The wind has an outward drift from the central part of the anticyclone, which is usually compensated for by descending air at higher levels. The descending motion dissolves the high and medium clouds. The anticyclone is therefore often a region of stable and fair weather. In winter, radiation fog frequently forms over land in the stagnant parts of the anticyclones.

The anticyclones do not always give fair weather. Fronts may sometimes penetrate far into the anticyclonic regions; over oceans and warm continents, convection may give showers. The weather is usually fair in anticyclones with increasing pressure, but the dissipating anticyclones may have unsettled or bad weather in places.

Vertical Extent of Cyclones and Anticyclones.—It is easy to show that the vertical extent of these pressure systems depends greatly on the air temperature. The equation of static equilibrium (see page 51) may be written as

$$-\frac{\Delta p}{\Delta z} = \rho g$$

Since the air density ρ increases with decreasing temperature, it follows that the pressure will decrease more rapidly along the vertical in colder air than in warmer air.

We consider first a warm anticyclone. The pressure is high at the earth's surface; and since it decreases slowly with elevation, it follows that a warm anticyclone (such as the subtropical anticyclone) will maintain itself in the upper atmosphere. Conversely, in a cold anticyclone the pressure is high at the earth's surface, but it decreases rapidly with increasing elevation. A cold anticyclone (such as the Siberian High in winter) will therefore be relatively shallow. Exactly in the same way, we find that a cold cyclone will maintain itself in the upper atmosphere, whereas a warm cyclone will be relatively shallow. It will be seen from Figs. 72 and 76 that the above rules hold. Thus, for example, the cold high over Siberia in winter is so shallow that it is not noticeable at the 6000-ft. level, whereas the warmer subtropical anticyclones are maintained at upper levels.

CHAPTER XI

WEATHER ANALYSIS

The aim of weather analysis is to "diagnose" the weather situation in all three dimensions. The result of this analysis leads to a more or less complete picture of the physical state of the atmosphere. By analyzing the weather situation three or four times a day, the meteorologists obtain a picture of the displacement of the fronts, cyclones, air masses, etc., and the changes that they undergo while they travel. On the basis of this information, the meteorologists forecast the future displacements and developments.

The theory and the principles of weather analysis and forecasting form one of the most intricate branches of meteorology, and a complete discussion of them is beyond the scope of this brief outline. We can here only indicate roughly how the analysis is performed and how forecasts are made.

Observations and Symbols.—Simultaneous observations of the state of the atmosphere are made at a large number of reporting stations (see page 10) on land and also in some selected ships. Such observations are made four times a day[1] with auxiliary observations at more frequent intervals. The actual observations are transcribed into coded messages which are transmitted by radio, teletype, or other means of telecommunication.

In order to facilitate the analysis of the vast mass of data, it has been found convenient to plot the reports on maps of suitable scale and projection. Each reporting station is represented on the map by a small circle. The meteorological elements, as reported by the station in question, are plotted around the station circle. For convenience, some elements are plotted as numerals and others as symbols, on the principle that instrumental observations, such as pressure and temperature are plotted as numerals that indicate the corrected readings, whereas noninstrumental

[1] In the United States the standard observations are made at 1:30 and 7:30 A.M. and 1:30 and 7:30 P.M. E.S.T.

observations (such as clouds and weather) are plotted in symbols. A system of convenient symbols has been adopted for international use by the International Meteorological Organization. The basic symbols of this system are shown in Fig. 106.

The amount of cloudiness is indicated by filling the station circle in an amount proportional to the cloud cover. The cloud symbols are made to indicate the characteristic features of the cloud forms that they represent. Thus, clouds of the cumulus

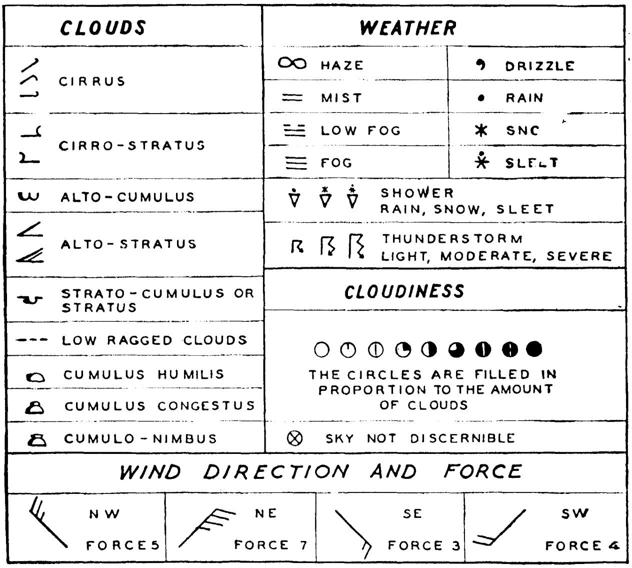


FIG. 106.—International symbols for use on weather charts.

family are represented by dome-shaped symbols, stratiform clouds by horizontal strokes, undulated clouds by an undulated curve, frontal clouds by a horizontal stroke indicating the stratiform structure and a sloping stroke resembling the inclination of frontal surfaces, and cirrus by feathery strokes.

The principal weather symbols are composed of horizontal strokes for fog, commas for drizzle, dots for rain, stars for snow, etc. The intensity and the continuous character of precipitation are indicated as follows: Two equal symbols side by side indicate "continuous"; two equal symbols, one above the other, indicate "increased intensity"; two unequal symbols, one above the other, indicate "coexistence" of the weather phenomena that the symbols represent. Thus, for example, one rain symbol alone means light and intermittent rain; two rain symbols side by side mean light continuous rain; two rain symbols, one above the other, mean intermittent rain of medium intensity, and three rain symbols arranged in a triangle mean continuous rain of medium intensity; three rain symbols above one another indicate heavy intermittent rain; and four symbols arranged as ⁖ indicate heavy and continuous rain. The same principle applies to the symbols for drizzle, snow, and sleet. If the precipitation is of a showery character (instability), a triangle with the apex downward is added to the precipitation symbol.

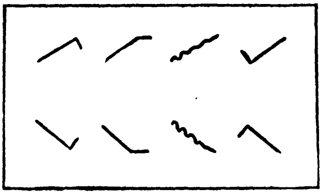


FIG. 107.—Each symbol represents the principal trend of the barograph trace during the last 3 hr. No symbol is used for steady rise or steady fall.

The wind direction is indicated by an arrow "flying with the wind" and ending on the station circle (see Fig. 106). The wind force is indicated by "feathers," so that each whole feather represents two points of the Beaufort scale of wind force (see Table V), and each half feather represents one point.

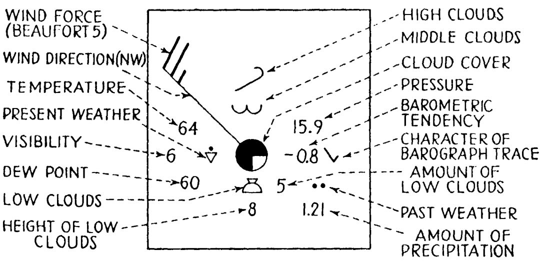


FIG. 108.—International plotting model for land stations. A similar model is used for ship stations.

Symbols are also used to indicate the characteristic of the barograph trace (see Fig. 107).

In order to facilitate the interpretation of the weather maps, the various elements, or symbols, are plotted in a definite order around the station circle. Figure 108 shows the international plotting model, and Fig. 109 the plotting model used in the U.S. Weather Bureau.

After the completion of the plotting, the analysis commences; and, again, it is necessary to use symbols in order to indicate the results of the analysis. On ordinary working maps, it is most convenient to use colored pencils to indicate the analysis, but on printed maps other symbols have to be used. Table IX shows the symbols recommended by the International Meteorological Organization.

Drawing of Isobars.—Throughout the history of weather forecasting, it has been recognized that the distribution of atmos-

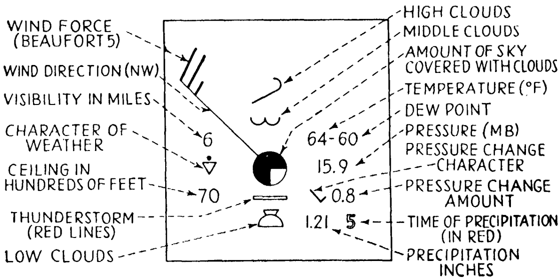


FIG. 109.—Plotting model used in the U. S. Weather Bureau.

pheric pressure is profoundly related to the weather phenomena and the changes in the weather situation. The early methods of weather forecasting were almost entirely based on studies of types of isobaric picture. In order to benefit fully from the modern methods based on the study of air masses and fronts, it is necessary to draw the isobars with meticulous care.

It should be emphasized at the outset that a satisfactory analysis can be based only on a succession of weather charts, not on a single chart. Only a series of maps will show the life history of the air masses, fronts, cyclones, etc.

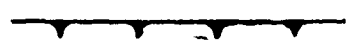
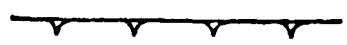
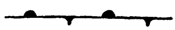
It is usually convenient to sketch in preliminary isobars, which may be revised later as the study of the air masses, fronts, etc., progresses. In drawing isobars the following principles should be borne in mind.

1. *The isobars should have such direction and mutual distance that they agree with the relation between wind and pressure gradient as explained in earlier sections.* The application of this principle

is shown in Fig. 110 where the isobars are so spaced that the wind force is proportional to the pressure gradient and the wind direction deviates from the isobars in the normal manner.

2. *In representing the large-scale movement of the air, simple isobars are much more probable than complicated isobars.* The

TABLE IX.—INTERNATIONAL SYMBOLS FOR RESULTS OF ANALYSIS

Phenomenon	On working charts	On printed charts	Direction of motion of fronts
Cold front at the ground.	Continuous blue line		↓
Cold front above the ground.	Broken blue line		↓
Warm front at the ground.	Continuous red line		↑
Warm front above the ground.	Broken red line		↑
Occluded front at the ground.	Continuous purple line		↑
Occluded front above the ground.	Broken purple line		↑
Quasi-stationary front.	Red-and-blue line		No motion
Areas of fog.	Continuous yellow area	Distributed fog symbols	
Areas of tropical air.	Continuous red area	Dotted area	
Areas of continuous precipitation.	Continuous green area	Hatched area	

application of this principle is shown in Fig. 111. In Fig. 111*a* the isobars are drawn strictly according to the observations. Now, the actual observations may, and frequently do, contain minor errors that cause irregularities in the isobars. Irregularities that do not show any systematic arrangement are likely to be due to errors; they should therefore be eliminated through smoothing. On the other hand, irregularities that show a systematic arrangement, such as troughs or wedges, should be regarded as real.

Let us imagine that the isobars in Fig. 111*a* run either around a center of low pressure (*i.e.*, 990, 995, and 1000) or around a center

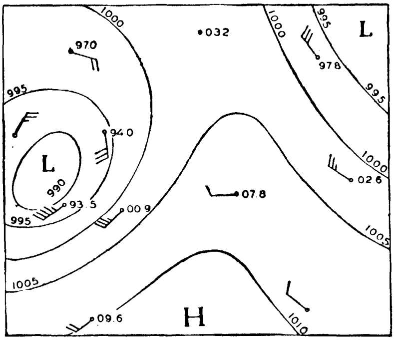


FIG. 110.—The isobars should be drawn so as to fit with the pressure and the direction and force of the winds.

of high pressure (*i.e.*, 1020, 1025, and 1030). The small wavelike irregularities do not show any systematic arrangement; they should therefore be eliminated by smoothing. Two troughlike or wedgelike irregularities then remain, one to the lower right of the center and the other to the lower left of the center. As these irregularities are systematically arranged and depend on more

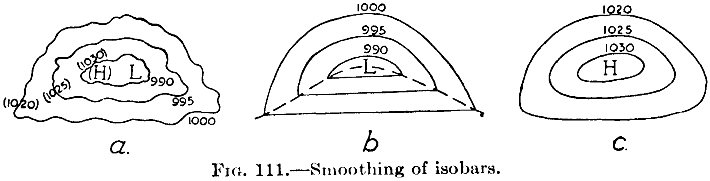


FIG. 111.—Smoothing of isobars.

than one station, they should be regarded as real. Figure 111*b* shows the smoothed isobars around a cyclonic center. It will be seen that these isobars have a marked resemblance to the isobars around a young wave cyclone. Whether or not there is a front in the pressure trough must be decided by applying the frontal characteristics described in a previous chapter. If it is evident that there is no front, the isobars should be drawn without kinks.

We imagine next that the isobars in Fig. 111*a* belong to a center of high pressure. The smoothed isobars would then be as shown in Fig. 111*c*. No kinks should be drawn there, for the kinks would then point from high to low pressure, which is impossible.

3. *The isobars in the vicinity of fronts should be drawn so as to bring out the frontal discontinuity in the horizontal pressure gradient.* The application of this principle is shown in Fig. 112. The observations seem to indicate that a warm front is present between the two stations. Use principle 1, and draw the isobars in the vicinity of the two stations; prolong the isobars thus drawn to intersection as shown by the broken lines. Since the front must coincide with the kinks in the isobars, it follows that the position of the front is determined by the intersection of the isobars. Through consistent drawing of isobars, the position of fronts can, in many cases, be determined with great accuracy.

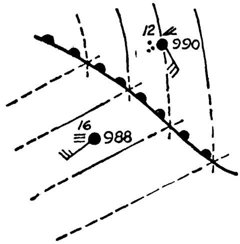


FIG. 112.—Examples showing how to draw isobars in the vicinity of a front and how to find the position of a front.

4. *The isobars should be drawn first in those areas where the analysis is simplest and should then be prolonged into areas where the solution is more difficult.* Obviously, it is easier to draw isobars over land areas where observations are numerous than over ocean areas where observations are sparse. Other conditions being equal, it is easier to draw isobars in regions where the winds are strong than in regions where the winds are light and variable. This is due to the fact that the light winds are more subject to local influences than are strong winds. Thus, in Fig. 113 the isobars should first be drawn in the vicinity of the stations *A*, *E*, *D*, and *B*. When this has been done, the isobars should be connected as shown by the broken lines.

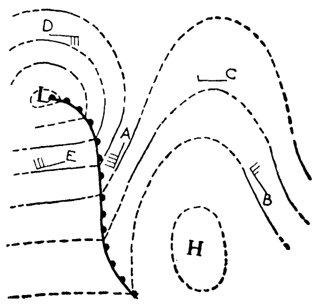


FIG. 113.—Full lines indicate the isobars that should be drawn first; broken lines indicate the isobars that result from connecting the different areas.

5. *The isobars should be drawn so as to agree logically with the previous map.* Thus, if the previous map showed a young deepening cyclone, the following map should be drawn on the assumption that the center has moved a reasonable distance and that the pressure in the center has decreased. It is important to apply this principle to ocean areas when observations are sparse.

Drawing of Isallobars.—The change in atmospheric pressure during the 3 hr. preceding the observation is called the *barometric tendency.* When the tendencies have been plotted on the map, lines may be drawn through the points that have the same tendency. Such lines are called *isallobars.* The isallobars, therefore, represent the pressure changes in the same way as the isobars represent pressure. The isobars are drawn for every 5 mb., but the isallobars are drawn for each millibar. In the same way as the pressure gradient is perpendicular to the isobars and inversely proportional to the distance between the isobars, the *isallobaric gradient* is perpendicular to the isallobars and inversely proportional to the distance between the isallobars.

The importance of the isallobars lies in the fact that they indicate how the pressure systems are moving and changing in structure. A few examples will be discussed later.

Analysis of Weather Charts.—The aim of the analysis is to obtain a picture of the three-dimensional structure of the atmosphere. If upper-air observations are available in a sufficient number from a sufficiently large area, the analysis is greatly simplified. In many cases, it is necessary to analyze the weather situation without the aid of observations from the free atmosphere. In such cases the problem is rather to evaluate the conditions at the earth's surface and to draw conclusions as to the structure of the free atmosphere. In doing this, the analyst uses those of the surface observations which depend mainly on the conditions in the free atmosphere. Thus, for example, the clouds and the various forms of precipitation as reported by the observers indicate the condition of the free atmosphere, and from these observations the analyst will be able to draw conclusions as to the stability conditions of the upper air.

1. *The first step* in the analysis is, therefore, to examine the forms of clouds and precipitation, to indicate the fronts between the various air masses, and also to indicate whether the air masses in question are stable or unstable. Thus, clouds of the cumulus

family and showery or squally precipitation are indicative of instability; fog, clouds of the stratus type, and precipitation of the drizzle type are indicative of stability; cirro-stratus, alto-stratus, and nimbo-stratus and extensive areas of even precipitation are indicative of fronts.

2. *The second step* is to identify the various air masses, fronts, and cyclones with those indicated on the preceding map. The fronts and air masses must, of course, move with the speed of the prevailing winds, and their positions on the present map must agree logically with their positions on the previous map and the air motion in the interval between the maps. Thus, the analysis must conform strictly with the principle of *historical sequence.* This is one of the fundamental principles in weather analysis.

3. *The Third Step.*—The time has now come to draw the isobars. It is preferable to draw all the isobars in one air mass, and then all the isobars in another air mass, and so on. While drawing the isobars in this way, the analyst studies the properties of each air mass. The isobars should, therefore, be drawn "from front to front," and the isobars in the vicinity of the fronts should be adjusted as shown in Fig. 112 in order to locate the fronts as accurately as possible.

4. *The fourth step* in the analysis will usually be to examine the air masses in detail in order to clarify the following points: (*a*) the origin of the air mass in question, (*b*) whether it travels toward warmer or colder regions, (*c*) whether the diurnal heating or cooling will change the stability conditions, etc.

5. *The fifth step* in the analysis will usually be to examine in all possible detail the conditions along the fronts, the type of front (whether cold, warm, or occluded, etc.) all frontal characteristics described in earlier sections being used, and also the previous map for identification of the various fronts.

6. *The Sixth Step.*—Although the barometric tendencies should be used to locate fronts, it is convenient to leave the actual drawing of the isallobars till the fronts and isobars have been drawn in a final form.

7. *The seventh step* in the analysis is to examine the regions where the wind distribution is favorable for the formation of new fronts. If the winds converge and if there is a sufficient temperature gradient, fronts will form as explained in an earlier section.

When this program has been completed, the map should be colored, the symbols listed in Table IX being used.

If upper-air observations are available, the analysis of the surface map should be compared with the adiabatic charts and other upper-air charts in order to ascertain that the analysis of the surface map is in agreement with the upper-air data.

Examples.—Figure 114*A* shows a highly simplified weather map. In this particular case, the difficulty in the way of obtaining a satisfactory analysis lies in the fact that the reporting stations are widely scattered; but by performing the analysis step-wise in the logical sequence, it is quite easy to obtain a correct analysis.

Step 1.—The clouds indicate that the southeastern and eastern portions of the map are occupied by a stable air mass, whereas the northwestern portion is occupied by an unstable mass. The prevailing winds indicate the presence of two air masses of widely different life history. The forms of precipitation confirm this: there are instability showers north of the British Isles and fog or drizzle over the North Sea. In addition, there is evidence of a frontal rain area from the Shetland Islands, across England down to the west of the English Channel. The two air masses are therefore separated by a distinct front. Although the position of this front may be sketched on the map, its character and exact position will have to be determined in greater detail as the analysis proceeds.

Step 2.—The time has now come to compare the present map with the preceding one (which is not reproduced here). The suggested front should be identified with the front on the preceding map, and it should be ascertained that its displacement from the preceding to the present map is in agreement with the principle of historical sequence.

Step 3.—Draw preliminary isobars, following the rules described in a previous paragraph. First draw the isobars in the air mass that has the strongest winds. Apply the rules for locating the position of the fronts (see Fig. 112). In addition, the barometric tendencies and the characteristics of the barograph traces should be used in order to find out approximately when the front passed the stations in the rear of the front. If there are no kinks in the barograph trace, the front must have passed more than 3 hr. before the hour of observation; if there are

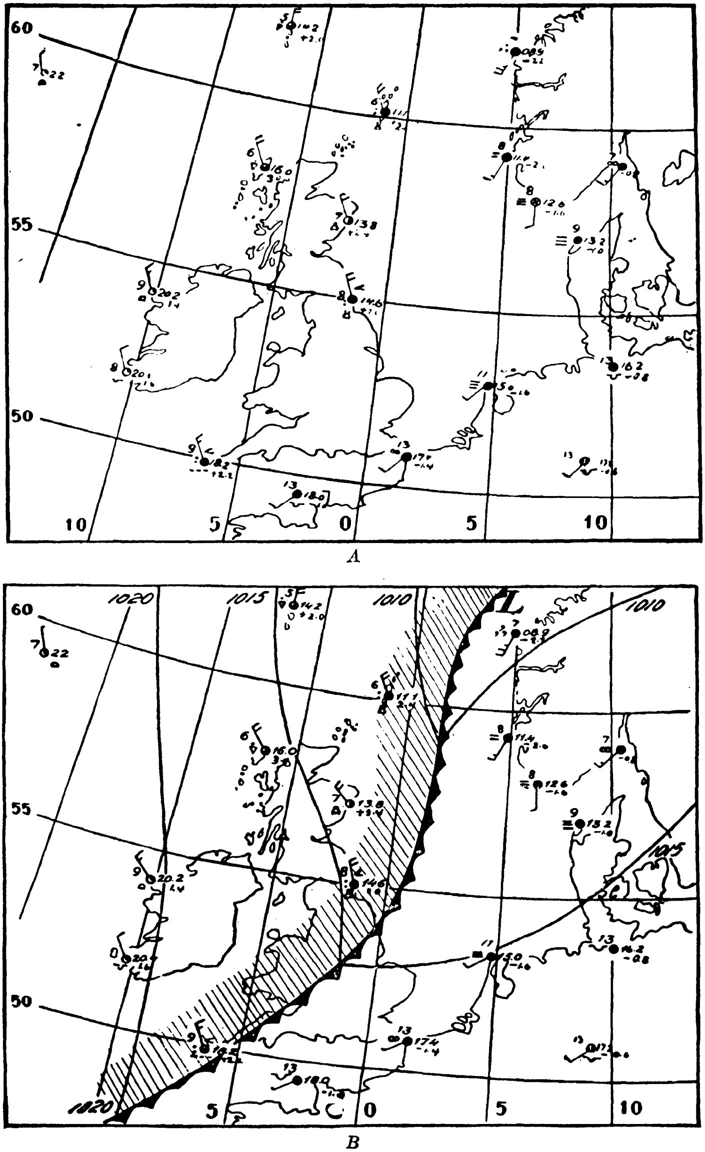


FIG. 114.—*A*, simplified weather chart for exercise; *B*, correct analysis.

frontal kinks with large positive tendencies, the front will have passed early in the 3-hr. interval; and if there are frontal kinks with negative tendencies, the front is likely to have passed in the late part of the tendency interval. This and the drawing of the isobars will do much to assist in finding the exact position of the front.

Step 4.—When the isobars have been drawn, a clearer picture of the life history of the air masses will be obtained. Comparison with the previous map (or maps) should now be made in order (1) to identify the air masses and (2) to ascertain whether the air masses are developing toward increased or decreased stability. In the present case, which is a simple one, it is evident that the air in the rear of the front is of polar origin, whereas the other mass most probably is of tropical origin.

The warm mass streams slowly northward and will be cooled from below. Furthermore, since the wind speed is low, there will not be much vertical mixing, and the mass is likely to preserve the temperature inversion at the ground that is suggested by the numerous fog observations.

The unstable polar air streams southward and will therefore be heated from below; this will tend to increase the lapse rate, and convective clouds and showers may result. However, the absence of showers and towering cumulus in the western portion of the map seems to suggest that the heating from below is counteracted by subsidence, with the result that not much change in lapse rate will result. The fact that the stations on the west coast of Ireland report strato-cumulus indicates that this air is relatively stable.

Step 5.—A detailed examination of the front shows that it is a cold front; the rain area is in the rear of the front, where the temperatures are lower than in advance of it; furthermore, the tendencies show rising pressure in the rear and falling pressure in advance of the front.

Step 6.—Draw isallobars in both air masses.

Step 7.—On account of the pronounced convergence toward the front, the front will maintain itself; but there is no other region on the map within which there is sufficient convergence to cause new fronts to form.

Figure 115 shows a highly interesting and instructive example of an apparently intricate situation that, subjected to a rational analysis, becomes very simple. This example, which is offered as an exercise for the student, should be analyzed as described earlier. In addition, study of the distribution of the isallobars is recommended, in order to find whether the wind normal to the front will increase or decrease.

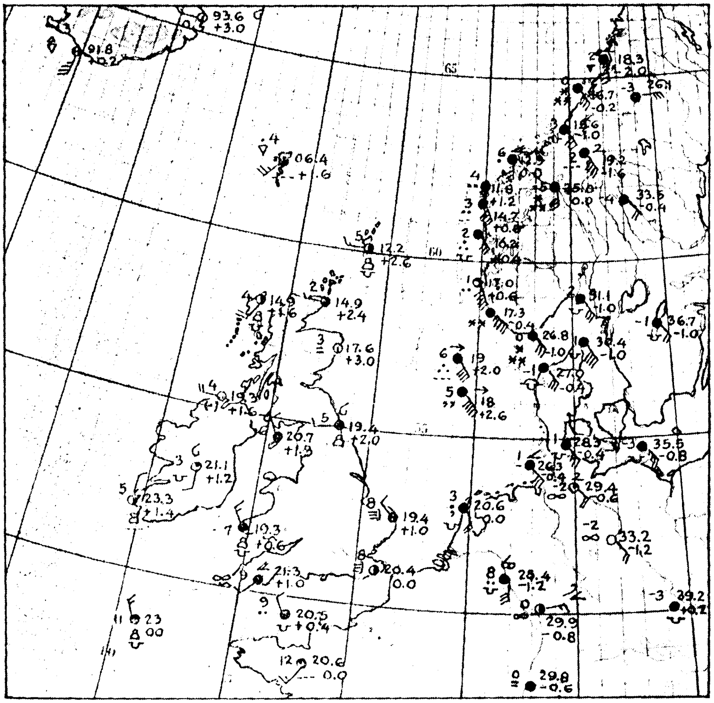


FIG. 115*A*.—Simplified weather map, Jan. 13, 1937, 6 P.M., G.M.T.

To guide the student in the analysis, it may be stated that the front over the North Sea must be essentially a warm front; this is suggested by the area of precipitation and the clouds. However, the warm air in the rear of the front is definitely not of tropical origin; for its temperatures are too low, and it shows indications of instability. The front is, therefore, most probably a warm-front type occlusion.

Isentropic Analysis.—In recent years, through the work of Rossby and his collaborators, a new method of analysis of the conditions in the free atmosphere has come into general use. This method is based on the fact that the temperature variations in the free atmosphere are mainly adiabatic. In a dry-adiabatic process the potential temperature stays constant, while the air moves upward or downward. Thus, a unit of air that moves

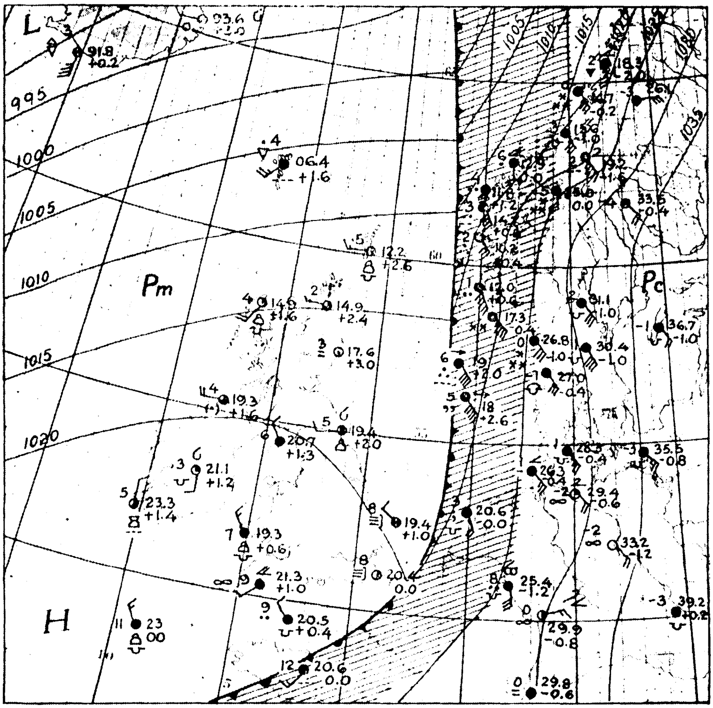


Fig. 115*B*.—Correct analysis of map, Fig. 115*A*.

up or down will remain in a surface of constant potential temperature.

In a dry-adiabatic process, there is no heat added to or withdrawn from the air, and then not only the potential temperature but also the entropy of the air stays constant. A surface of constant potential temperature is therefore also a surface of constant entropy. Now, in order to identify the air from one map to the next, we can study the conditions in suitable isentropic surfaces, which are so high in the atmosphere that they are not appreciably affected by the heating and cooling of the earth's surface.

In order to identify the air masses, we must have one more element that does not vary adiabatically. Such an element is the specific humidity. Therefore, by plotting the specific-humidity lines on an isentropic map and by studying the movement of these lines from one map to the next, we obtain a picture of the displacements of the air masses along isentropic surfaces. An example of an isentropic chart will be given in connection with a weather situation to be discussed later.

CHAPTER XII

WEATHER FORECASTING

It should be emphasized at the outset that there is no distinct line between the process of analysis and the process of forecasting. Almost every step in the analysis consists in considering the physical properties of the air in relation to space and time. The analysis, therefore, leads directly to a more or less vague conception of what will happen in the near future. When the analyses have been completed, the remaining problems may be formulated as follows:

1. To determine the movement of the pressure systems, fronts, etc., during the forecasting period.
2. To determine the changes in intensity that the pressure systems, fronts, etc., will undergo during the forecasting period.
3. To determine the changes in the physical properties of the air masses during the forecasting period.

Thus, the forecasting procedure is mainly concerned with the extrapolation of the future conditions on the basis of the present and past conditions. In actual practice, the forecasts are worked out on the basis of scientific principles as well as of years of experience. In this chapter, we can indicate only in brief outline the principles that are commonly used in forecasting the weather for 24 to 36 hr. in advance.

The Path Method.—The simplest method of determining the future movement of the pressure systems consists in a freehand extrapolation of their paths. Let the points 1, 2, 3, and 4 in Fig. 116*a* represent four consecutive positions of, say, a cyclone center at equal intervals of time. These positions show that the center has moved along a straight path with a constant speed. If, now, the distribution of pressure and pressure tendencies around the center in position 4 is approximately the same as around the center when it was in position 3, then we should be justified in assuming that the center would maintain its direction and speed. Consequently, after the end of the next time interval, its position would be as indicated by the point x.

In a similar manner, Fig. 116*b* shows a case where the center has moved on a straight path with decreasing speed. As a first approximation, we may assume that the center would move to the position x during the coming interval.

Figure 116*c* shows a center that has moved on a curved path with decreasing speed. Experience as well as theoretical considerations show that the speed of a pressure center decreases as the curvature of the path increases. If the path straightens again

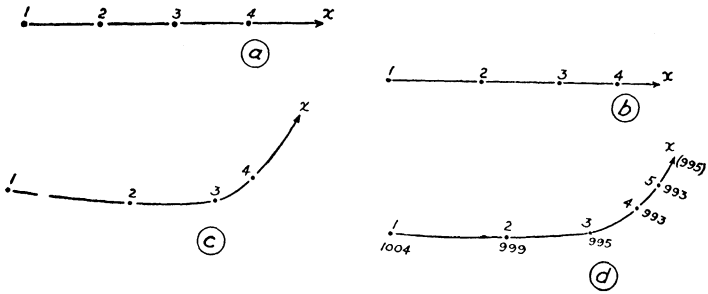


FIG. 116.—Illustrating the path method. The points 1, 2, 3, and 4 indicate consecutive positions of a pressure center, and x indicates the probable position on the next map.

after position 4 is reached, the speed is likely to increase. This increase in speed is not directly inherent in the previous path and must be determined by means of other methods.

Finally, Fig. 116*d* shows a center that moves on a curved path with decreasing speed. In addition, the pressure in the center is plotted along each position mark. It will be seen that the center deepened from 1004 to 993 mb. and that the rate of deepening has decreased. The logical forecast would then be that the center would move to the position x and that the pressure in the center would increase to about 995 mb. during the coming interval.

In a similar manner, the path method may be applied to anticyclonic centers, troughs, wedges, and fronts. The position of the trough, wedge line, or front would then be marked on the map by curves; and, from the displacement of these curves, the future position can be estimated.

The path method usually gives satisfactory results when maps are available at frequent intervals so that the paths can be studied in detail. The weakness of the path method lies in the fact that it fails when the paths of the pressure systems are subject to rapid changes.

The above discussion may be supplemented by the following forecasting rules:

Rule 1.—Cyclone and anticyclone centers will, in general, continue to move with the velocity and acceleration that they have had during the preceding 12 hr.

Rule 2.—A cyclone center that moves toward a stationary anticyclone is retarded and the path tends to curve northward until it becomes parallel to the isobars around the anticyclone.

Rule 3.—Warm-sector cyclones move in the direction of the current in the warm sector. They usually have straight paths, whereas old occluded cyclones usually have paths that are curved northward.

The Geostrophic-wind Method.—It was shown in the chapter on winds that the geostrophic wind is a good approximation to the actual wind at the top of the friction layer (approximately 3000 ft. above the ground). Thus, the geostrophic wind is more significant for the displacement of air masses than is the actual wind at the earth's surface. Since a front separates a colder air mass from a warmer one, it follows that a front must move in conformity with the air currents on either side of it. If there were no friction and if there were no vertical velocity in the air, then the velocity of a front would be simply equal to the wind velocity. Because of friction and the vertical velocity along frontal surfaces, the fronts tend to move somewhat more slowly than the geostrophic wind. Experience as well as theoretical considerations show that the following rules hold:

Rule 4.—Warm fronts move with a speed of about 60 to 80 per cent of the geostrophic wind.

Rule 5.—Cold fronts move with a speed of about 70 to 90 per cent of the geostrophic wind. Occasionally, the speed may exceed 90 per cent.

Rule 6.—Occluded fronts move as a warm or a cold front, depending on whether they are of the cold front or of the warm front type.

Rule 7.—A cyclone center moves with approximately the same speed as a warm front and somewhat more slowly than a cold front.

Rules 4 to 7 are readily applied to weather maps. The geostrophic wind depends on the distance between neighboring isobars, as shown on page 104. Table X gives the geostrophic displacement during 6 hr. for various spacing of the isobars and various latitudes. It should be noted that the geostrophic

TABLE X.—DISTANCE BETWEEN 5-MB. ISOBARS (MILES) GEOSTROPHIC WIND (MILES PER HOUR), AND GEOSTROPHIC DISPLACEMENT (MILES) IN 6 HR.

Distance between 5-mb. isobars	70°N. or S.		60°N. or S.		50°N. or S.		40°N. or S.	
	Geostrophic wind	Travel in 6 hr.	Geostrophic wind	Travel in 6 hr.	Geostrophic wind	Travel in 6 hr.	Geostrophic wind	Travel in 6 hr.
50	85	510	92	550	104	630	127	760
75	57	340	61	370	69	420	81	490
100	42	250	46	280	52	310	61	370
125	34	200	37	220	41	250	48	290
150	28	170	31	180	35	210	41	240
175	24	140	26	160	30	180	35	210
200	21	120	23	140	26	160	30	180
250	17	100	18	110	21	130	24	140
300	14	85	16	90	18	100	20	120

balance is less pronounced in low latitudes. The above rules are applicable only in middle and high latitudes, say north of 35°N.

The Tendency Method.—The displacement of pressure systems may be computed from the isobars and isallobars. To indicate how this is done, we consider a trough of low pressure as shown in Fig. 117*A*. The full lines are isobars and the broken lines are isallobars. If the barometric tendencies were uniform over the area covered by the trough, the trough would not move. For the trough to move, it is necessary that the tendencies in the rear of the trough be larger than in advance of it. The trough in Fig. 117*A* will then move in the direction from rising tendencies to falling tendencies, or the trough would advance in the direction of the *isallobaric gradient.*

Let us next draw an x-axis perpendicular to the trough line. On graph paper, we plot the pressure distribution along the x-axis

(Fig. 117*B*) and also the distribution of barometric tendencies along the x-axis (Fig. 117*C*). Thus, Fig. 117*B* gives us the "pressure profile," and Fig. 117*C* the "tendency profile" along the x-axis.

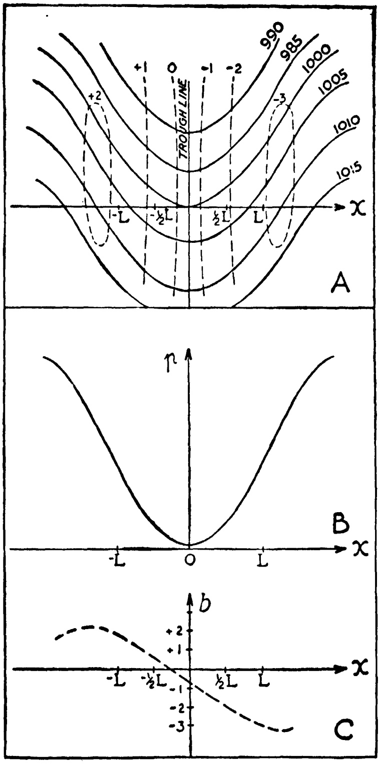


FIG. 117.—*A*, isobars (full lines) and isallobars (broken lines) in the vicinity of a trough of low pressure; *B*, pressure profile along the x-axis; *C*, tendency profile along the x-axis.

The velocity with which the trough line moves along the x-axis depends on two factors, *viz:*

1. The slope of the tendency profile, or, what is the same, the magnitude of the isallobaric gradient along the x-axis. Obviously, the larger the isallobaric gradient, the faster will the trough move.

2. The curvature of the pressure profile. Obviously, a pronounced trough has a pressure profile that is more curved than a flat trough. The more curved the pressure profile, the less effective is the isallobaric gradient in moving the trough.

Thus, the velocity V_x of a pressure trough along the x-axis may be expressed in the following formula:

$$V_x = \frac{\text{isallobaric gradient along the } x\text{-axis}}{\text{curvature of the pressure profile along the } x\text{-axis}}$$

If we consider a wedge of high pressure instead of a trough, we obtain the same result, with the exception that a wedge will move

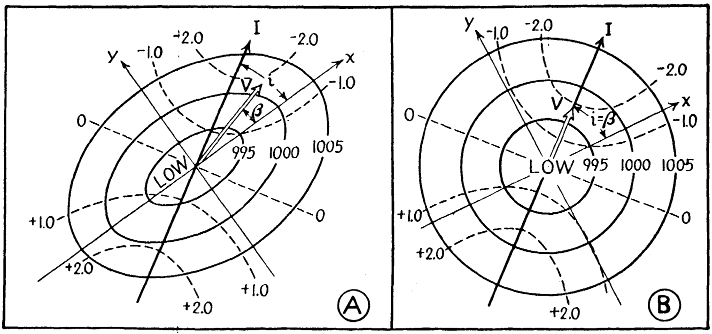


FIG. 118.—Showing the velocity in relation to the isallobaric gradient. *A*, for elliptical centers; *B*, for centers with circular isobars.

in the direction from falling to rising pressure. From the above formula, we obtain the following forecasting rules:

Rule 8.—Troughs move in the direction of the isallobaric gradient (*i.e.*, from rising to falling pressure), and wedges move in the opposite direction.

Rule 9.—The speed of the trough or the wedge is directly proportional to the isallobaric gradient and inversely proportional to the curvature of the pressure profile.

In general the following rule will hold:

Rule 10.—Troughs and wedges whose pressure profiles have great curvature (*i.e.*, pronounced troughs and wedges) proceed slowly, whereas troughs and wedges whose pressure profiles are slightly curved move with a speed that may vary within wide limits, depending on the isallobaric gradient.

We consider next a center of low pressure as shown in Fig. 118A. Draw a system of coordinates through the pressure center with the x-axis along the longest symmetry axis and the y-axis along the shortest symmetry axis. Each of these axes may be regarded as a trough line. The trough line coinciding with the y-axis will have a velocity V_x along the x-axis, and the trough line that coincides with the x-axis will have a velocity V_y along the y-axis. The cyclone center will then at any moment be determined by the intersection between these two trough lines. Consequently, the velocity of the cyclone center is determined by

$$V_x = \frac{\text{isallobaric gradient along the } x\text{-axis}}{\text{curvature of the pressure profile along the } x\text{-axis}}$$

$$V_y = \frac{\text{isallobaric gradient along the } y\text{-axis}}{\text{curvature of the pressure profile along the } y\text{-axis}}$$

Let I_x and I_y indicate the components of the isallobaric gradient along the coordinate axes, and let C_x and C_y denote the curvature of the pressure profiles along the same axes. Then,

$$(1) \qquad V_x = \frac{I_x}{C_x} \quad \text{and} \quad V_y = \frac{I_y}{C_y}$$

and the velocity **V** of the center is the resultant of these two components.

The angle β between the x-axis and the velocity **V** is given by

$$(2) \qquad \tan \beta = \frac{V_y}{V_x}$$

and the angle i between the x-axis and the isallobaric gradient (see Fig. 118A) is given by

$$(3) \qquad \tan i = \frac{I_y}{I_x}$$

Thus, the direction in which the center is moving is determined by the following formula:

$$\tan \beta = \frac{V_y}{V_x} = \frac{I_y}{I_x} \cdot \frac{C_x}{C_y}$$

or

$$(4) \qquad \tan \beta = \tan i \, \frac{C_x}{C_y}$$

It will now be seen that if the center is surrounded by elliptica isobars (as in Fig. 118A) $C_y > C_x$, with the result that

(5) $$\tan \beta < \tan i$$

If, however, the center is surrounded by circular or nearl circular isobars (as in Fig. 118B), then $C_x = C_y$ and

(6) $$\tan \beta = \tan i$$

If we consider an anticyclone instead of a cyclone, we obtai the same results, with the exception that the anticyclonic cente moves in the general direction from falling to rising pressure.

The above discussion may be summarized in the following fore casting rules:

Rule 11.—Circular (or nearly circular) cyclonic centers move i the direction of the isallobaric gradient, whereas anticycloni centers move in the opposite direction. The speed of the cente is directly proportional to the isallobaric gradient and inversel proportional to the curvature of the pressure profile.

Rule 12.—Cyclonic and anticyclonic centers whose pressur profiles have great curvature (pronounced centers) will mov slowly. When the pressure profiles are slightly curved (flat pres sure systems), the centers will move with a speed that varie within wide limits, depending on the isallobaric gradient.

Furthermore, from Eq. (4) we obtain

Rule 13.—Elliptical pressure centers move in a directio between the longest symmetry axis and the direction of th isallobaric gradient, and the more elongated the isobars are th closer is the velocity to the longest symmetry axis.

Rule 14.—Very oblong centers generally move along th longest symmetry axis, or very close to it.

Deepening and Filling.—The terms deepening and filling refe to the pressure variations in the individual pressure systems whil they move from one place to another. Thus, for example, a cyclone center is said to *deepen* when the pressure in the cente decreases with time and to *fill* when the pressure in the cente increases.

To obtain an idea of what deepening and filling mean, we con sider a cyclone that travels over an ocean. A number of island (fixed stations) furnished with barographs would then record th pressure changes associated with the cyclone. The barometri tendencies recorded by these stations would be partly due to the travel of the pressure system and partly due to the deepening or filling of the pressure system. Thus, if b_o denotes the observed tendency at a fixed station, we may write

$$b_o = b_m + b_d$$

where b_m is the part of the tendency that is due to the movement and b_d that due to deepening or filling.

Let us next suppose that a number of ships evenly distributed around the cyclone center sail with a speed and direction exactly equal to the speed and direction of the cyclone center. If the

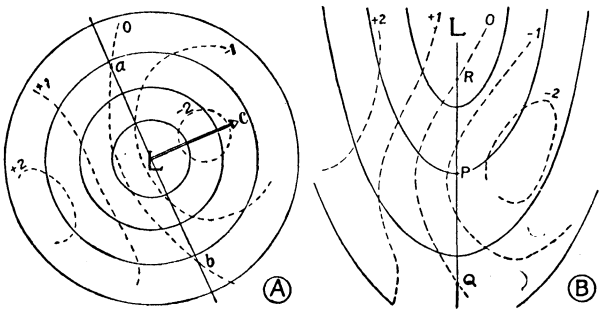


FIG. 119.—*A*, cyclone with deepening in the center; *B*, trough with deepening in the vicinity of *P*.

ships were furnished with barographs, they would record nothing but the part of the pressure change that is due to deepening or filling.

Now, in the pressure center there is no pressure gradient. Consequently, the movement of the center does not influence the barometric tendency reported at a fixed station. Therefore, $b_m = 0$ and $b_o = b_d$ in the center.

Consider now Fig. 119*A* where the cyclone center moves in the direction of the arrow. The isallobars indicate the tendencies observed at fixed stations. The fact that the barometric tendency in the center is approximately -1.2 mb. indicates that the center is deepening at a rate of 1.2 mb. in 3 hr.

Along the line *ab* in Fig. 119*A*, the isobars are parallel to the direction in which the center moves. Along this line the movement of the pressure system does not influence the tendencies observed at fixed stations. Therefore, $b_m = 0$ and $b_o = b_d$ along this line. It will now be seen that there is deepening along the line from b to a with a maximum of deepening in the center. North of the point a and south of the point b the tendency is positive and there is filling.

The above holds also for trough lines. Thus, in Fig. 119B, there is a maximum of deepening at P. At R and Q, there is neither deepening nor filling, and north of R and south of Q there is filling.

Exactly the same reasoning holds for anticyclones and wedges.

Let us next examine the conditions in a traveling warm-sector cyclone as shown in Fig. 120. Since the warm-center cyclone moves in the direction of the isobars in the warm sector, it follows that the movement of the pressure system does not influence the tendencies observed at fixed stations within the warm sector. Here, therefore, $b_m = 0$ and $b_o = b_d$. It will then be seen that the deepening at the peak of the warm sector in Fig. 120 is slightly greater than 2 mb. in 3 hr. In the southern part of the warm sector, the deepening is less than 1 mb. Since there is more deepening near the center than in the southern part of the warm sector, it follows that the pressure gradient will increase in the warm sector; the winds will therefore also increase.

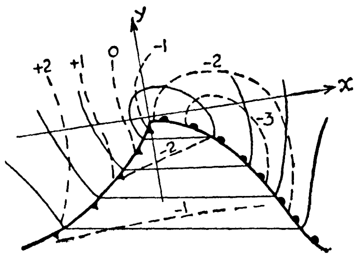


FIG. 120.—Wave cyclone with deepening in the warm sector.

The above discussion may be summarized in the following rules:

Rule 15.—A cyclone center deepens (or fills) with a speed that is equal to the barometric tendency in the center. The same also applies to anticyclonic centers.

Rule 16.—A cyclone center deepens when the zero isallobar is in the rear of the center, and it fills when the zero isallobar is in advance of the center. Anticyclonic centers obey the reverse rule.

Rule 17.—A trough deepens when the zero isallobar is in the rear of the trough line, and it fills when the zero isallobar is in advance of the trough line. Wedges obey the reverse rule.

Rule 18.—A warm-sector cyclone deepens with a speed that is equal to the tendency in the warm sector.

The Forecasting Procedure.—On account of the inadequacy of the observations and the shortcomings of our methods of analysis and prognostication, any forecast of future conditions is admittedly a hypothesis that, at least in parts, is based on assumptions. However, *the art of analysis and forecasting is to assume as little as possible and, as far as possible, to base the forecasts on conclusions drawn from actual observations.* To attain this, it is necessary to apply as many mutually independent methods as possible so as to obtain checks on the results derived.

The forecasting of future weather conditions is not a unique problem; it resolves itself into a number of partial problems that can be solved individually. It will then be found that certain of these partial problems can be attacked only after other problems have been solved. It is therefore important that the partial problems be attacked in the right sequence. Starting from a series of maps analyzed as shown in the previous paragraphs, the forecast may be approached in several steps, each dealing with the solution of one of the partial problems. These steps may be described in brief as follows:

Step 1.—The displacements of the pressure systems, fronts, etc., should be determined first. Usually, it will be possible to determine the displacements during 24 hr.; but, in uncertain or complicated cases, it is advisable to evaluate the displacements for 18 or 12 hr. only. It is always better to obtain an accurate determination of the displacements during a short interval of time than an inaccurate determination over a long interval of time.

The displacements should be determined numerically or by the qualitative forecasting rules. Whenever possible, the displacements should be determined by the three methods described in this chapter, *viz.*, the *tendency method*, the *geostrophic-wind method*, and the *path method.* These methods are largely independent of one another, and the results can thus be checked.

Step 2.—After having determined the displacements in a preliminary fashion, the deepening or filling of the various pressure systems should be examined as described above.

Step 3.—From the examination of the distribution of deepening and filling, evidence of increase or decrease in the intensity of preexisting systems (cyclones, anticyclones, troughs, wedges, fronts, etc.) will be found. In addition, evidence of the formation of new systems might be found. The third step in the forecasting procedure is therefore to determine whether or not new systems will form and what influence they will have on the weather situation.

Step 4.—We should now try to determine the position and the properties of the air mass (or masses) that is going to pass over the forecasting district during the forecasting period. This is, of course, a direct consequence of the travel of fronts and air masses. The properties of the air mass (or masses) should be examined in great detail. A detailed analysis of the clouds, hydrometeors, ascent curves (adiabatic charts), moisture patterns aloft (isentropic charts), and pressure charts aloft will reveal the conditions of the air mass in question.

Step 5.—An estimate should next be made of the changes in the physical properties of the air mass (or masses) that will occur by the time it arrives within the forecasting district. The possibilities for heating or cooling, depletion or supply of moisture, subsidence, etc., should be considered.

Step 6.—The modifications caused by local influences should next be considered—*e.g.*, the influences of mountain ranges, valleys, lakes, land and sea breezes, and other local effects.

Step 7.—When the foregoing points have been completed it is well to reexamine the weather charts in order to ascertain whether any phenomenon or alternative has been overlooked. In other words, the analyst should ask himself the question: What can upset the forecast?

Step 8.—The final point to consider is the wording of the forecast. The forecast should be as clear and unambiguous as possible, and it should express not only the weather conditions but also the degree of certainty of the prediction. For example, it may be quite certain that rain will occur, but somewhat uncertain whether the rain area will arrive in the morning or in the evening. In another case, it may be certain that a front will pass about noon, but not certain whether or not the front will cause precipitation while it passes. The forecasts should contain some indication of the certainty of the prediction, both as far as weather phenomena and the time of occurrence are concerned.

CHAPTER XIII

EXAMPLES OF WEATHER MAPS

In this chapter a few interesting weather situations will be discussed in order to demonstrate in outline the application of the methods of analysis and forecasting. The maps are plotted according to the international plotting model (Fig. 108), and the symbols used are shown in Figs. 106 and 107 and Table IX (page 168).

Example of Ocean Analysis.—In order to demonstrate the application of the principles of analysis and forecasting to ocean areas where observations are scanty, we turn to the weather situation for Feb. 12 to 14, 1937, over the North Atlantic. The maps (Figs. 121 to 125) are highly simplified so as to allow sufficient reduction. During this period the large-scale weather situation is characterized by

1. A large cold anticyclone over Siberia, Russia, and Scandinavia.
2. A large warm anticyclone between Spain and Bermuda.
3. A cold continental anticyclone over the North American continent.

Comparison with Fig. 72 will show that these anticyclones are in their normal positions; they constitute the normal wheels in the general circulation.

The cold air from the Arctic and north Canada streams southward and meets with the tropical maritime air from the Atlantic subtropical anticyclone along an extensive and pronounced polar front which is approximately in its normal position (*cf.* Fig. 88). A number of cyclone waves form along the polar front and go eastward, while they develop in the normal manner (see page 156).

Figure 121 shows the situation Feb. 12, 1:00 P.M. G.M.T. We shall first examine briefly the properties of the various air masses. In Newfoundland, the air temperature is about $-13°C$. Since the winds are strong and the air is heated from below, the reported temperatures are altogether representative. By passing from

Newfoundland to the ship indicated by *a*, the air temperature has increased to +1°C., but the air is still considerably colder than the sea surface.

The ships *b*, *c*, and *d* report air temperatures varying from 15 to 19°C., and the temperature of the air is about 1°C. higher than the temperature of the sea surface. This indicates that the air is of warm (subtropical) origin and that it is cooled from below. Normally, this would cause inversions to form in the lower layers; but since the winds are very strong, the vertical mixing is sufficiently intense to hinder the formation of appreciable inversions. As a result, none of the ships in this region reports stratus, fog, or drizzle.

The polar air that streams toward Florida is headed rapidly over the warm Gulf Stream. In spite of this, none of the ships in this area reports showers or squalls. This may be due to anticyclonic subsidence which, as far as stability is concerned, counteracts the heating from below.

The front between the two air masses is very pronounced. In spite of the scarcity of ship reports, the position of the front may be located with great accuracy by drawing the isobars as described on page 167. It will be seen that the temperature difference along the front varies considerably; it is greatest in the rear of the cyclone *B* where the polar air has had a minimum of travel over warmer water. In the eastern part of the Atlantic, the temperature difference along the front is considerably smaller. This is due to the fact that the polar air has traveled a long while over the ocean. However, as the maritime air proceeds eastward and meets with the cold air over the European continent, the front will increase in intensity.

Figure 121 gives an idea of the extent and intensity of the polar front in pronounced winter cases. The reader is advised to use the symbols given on pages 165 to 168 and study the maps in order to become familiar with the interpretation of the symbols. A complete discussion of the analysis and the forecasts concerning these maps has been published elsewhere.[1] We shall therefore point out only a few salient features.

It is readily seen from Fig. 121 that the wedge of high pressure over Scandinavia will remain almost stationary and build up

[1] PETTERSSEN, SVERRE, "Weather Analysis and Forecasting," McGraw-Hill Book Company, Inc., New York, 1940.

slightly. That this is so follows from rule 9 (page 184). Since the tendencies are almost uniform across the wedge, there cannot be much displacement. Likewise, since the tendencies on the wedge line are slightly positive, it follows from rule 15 (page 188) that the wedge will build up slightly.

We consider now the occluded front A. Using the gradient-wind method, we find that the front should move as indicated by the thin arrows. However, since the wedge farther to the east will remain stationary, the speed of the front will decrease. The appropriate displacement of the occluded front during the coming 18-hr. period is indicated by the heavy arrows.

The fact that the occluded front A is retarded is indicated also by the barometric tendencies. The pressure is rising near A and falling in front of the center B. The isobars in the rear of the front A will therefore rotate toward the south; as a result, the geostrophic wind normal to the front will decrease. The speed of the front will therefore decrease too.

Applying next the path method to the center near the east coast of Greenland, we find that the movement of the cyclone center is strongly retarded. The tendency method gives the same result. It will be seen that the barometric tendency in south Greenland is -1 mb. and only -2 mb. at Jan Mayen (in advance of the center). Since the isallobaric gradient is very small whereas the curvature of the pressure profile is large, the center will not move appreciably.

We next consider the cyclone B. Using the path method and the geostrophic-wind method, we find that the center will move during the coming 18-hr. period as indicated by the arrow. This cyclone has a wide warm sector, and the temperature contrasts are very large. The supply of energy is, therefore, very great, and a rapid development (deepening) is to be expected. Note that the barometric tendency at the ship d is $+0.4$. This does not indicate deepening (*cf.* rule 18, page 189). However, the ship sails eastward as indicated by the small arrow. Correcting the barometric tendency for the movement of the ship, we find that the correct tendency is about -3 mb., which is indicative of rapid deepening.

As the center B moves toward the northeast, the cold polar air from Canada will stream eastward in the rear of the cyclone.

Figure 122 shows the situation 6 hr. later. The reader is advised to analyze this map and apply the forecasting rules to obtain an idea of the development during the coming 12-hr. interval.

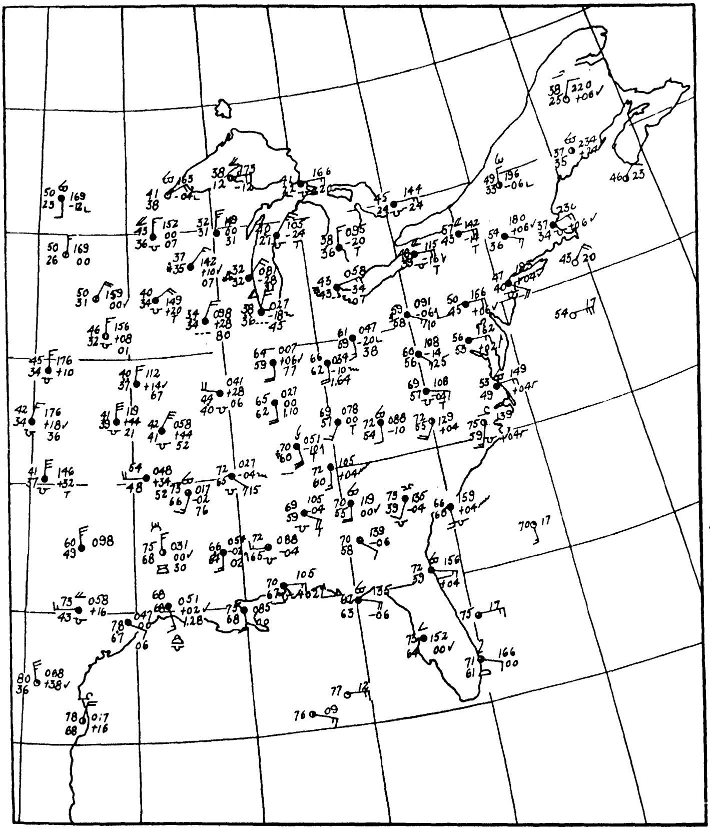


FIG. 126.—Unanalyzed weather map. Apr. 17, 1940, 7.30 P.M., E.S.T.

Figure 123 shows the situation 18 hr. after the first map (Fig. 121). It will be seen that the cyclones, fronts, and air masses have moved in accordance with the prediction.

The further development is seen in Figs. 124 and 125. On each of these maps the arrows indicate the forecast displacements.

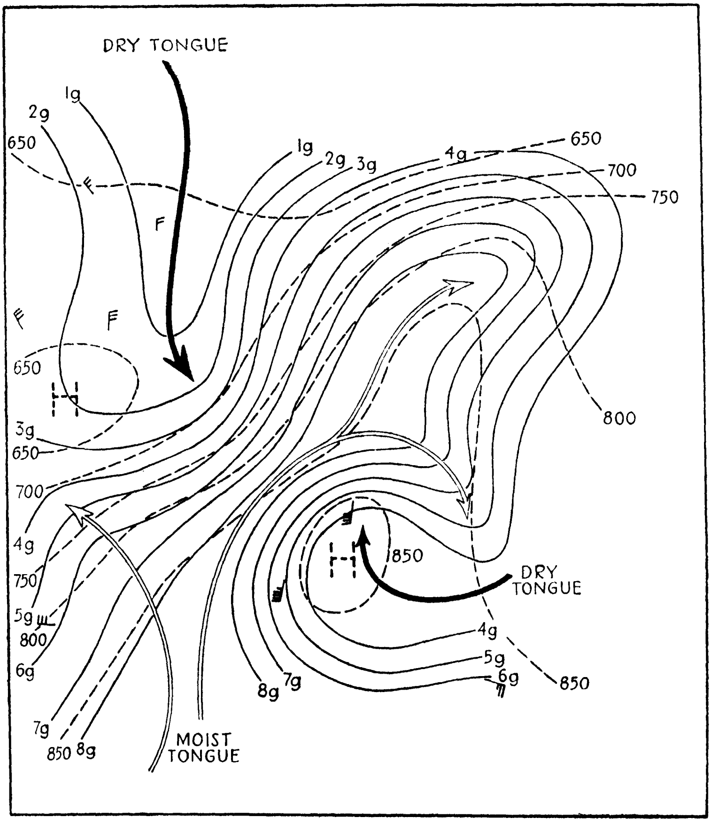


Fig. 128.—Isentropic map, Apr. 18, 1940, 1.30 A.M., E.S.T. Broken lines indicate the atmospheric pressure in the isentropic surface. The isentropic surface is high where the pressure is low and vice versa. Full lines represent specific humidity expressed in grams. Double-shaft arrows indicate moist tongues and single-shaft arrows indicate tongues of dry air.

Note the following features: (1) The occluded front *A* becomes stationary and dissolves as it approaches the cold continental anticyclone. (2) The cyclone *B* deepens and occludes as described on page 156. (3) The path of the advancing cyclone

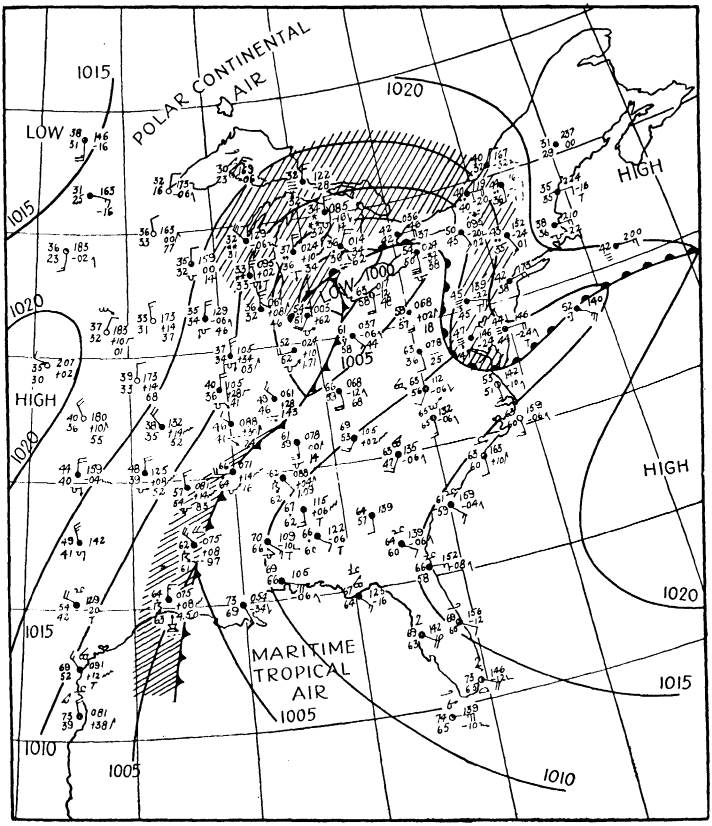


FIG. 127.—Weather map, Apr. 18, 1940, 1.30 A.M., E.S.T.

curves northward as it approaches the quasi-stationary anticyclone over Europe. As soon as the path has become parallel to the isobars around this anticyclone, the cyclone center resumes a straight path. (4) The southwest portion of the cold front associated with the cyclone *B* runs into a col where conditions are favorable for frontogenesis. While this happens, a well-developed rain area develops along the front.

Example of Three-dimensional Analysis.—We shall now give an example of a weather situation analyzed in all three dimensions. Figure 126 shows an unanalyzed map that is offered as an exercise for the reader. A broad mass of tropical maritime air is present over the southeastern part of North America. The polar front is, in this case, displaced far to the northwest of its normal position where the tropical maritime air meets with a mass of polar continental air. The disturbance has the shape of a young wave cyclone with its center to the southeast of Chicago. Apply the methods described in earlier sections, and complete the analysis.

Figure 127 shows the situation 6 hr. later. It will be seen that the center has moved in the direction of the warm-sector isobars. The width of the warm sector has decreased. In the rear of the center, two isobars cross the cold front. Here, the geostrophic wind normal to the front is strong, and the cold front will move with considerable speed. Farther to the south, the isobars are almost parallel to the cold front, which in this part of the map will move slowly. As a result of this, the cyclone will occlude only in the northern portion where the cold front moves quickly. Furthermore, the cold front in the southwestern portion of the map runs through a col. It will therefore be exposed to frontogenesis. The strong wind shear along this part of the front seems to indicate that a new wave will form. Figure 129 shows the development 6 hr. later. The cyclone has advanced farther to the northeast and occluded, and a new wave is beginning to form on the southwestern position of the front.

Figure 130 shows the structure of the free atmosphere in a cross section from Oakland, California, to Portland, Maine. The cold front reaches slightly above the 3 km. (10,000-ft.) level, and the cloud system has built up to about 5 km. (15,000 ft.).

It was mentioned in an earlier chapter that the air in the free atmosphere tends to stream in an isentropic surface. Figure 128 shows the isentropic map corresponding to Fig. 127. A tongue of moist air from the Gulf of Mexico streams up through the trough in the isentropic surface along the fronts. It is this tongue of moist air that feeds the frontal disturbance with moisture. In the cold air the isentropic surface is high and the air streams

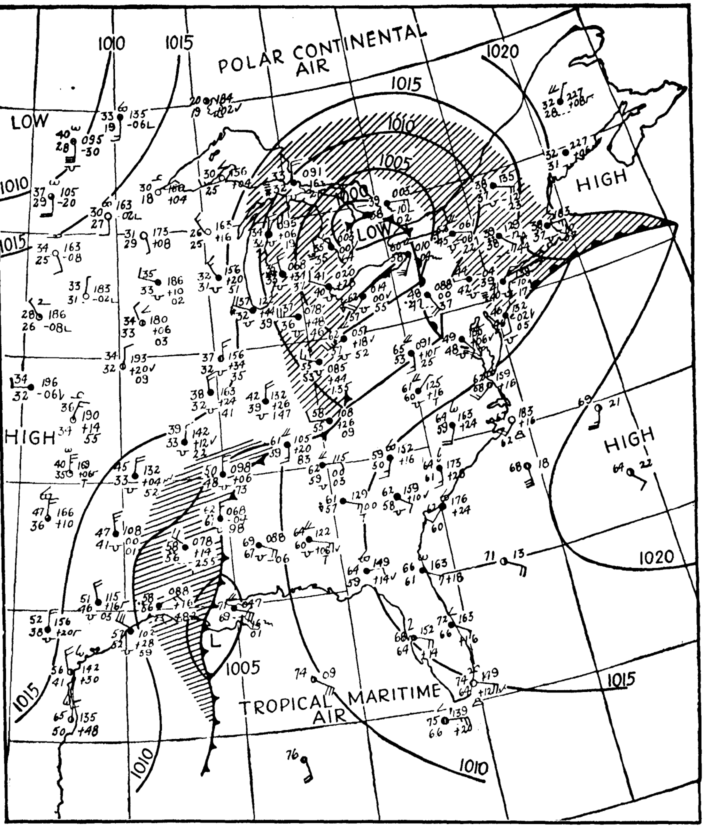


FIG. 129.—Weather map, Apr. 18, 1940, 7.30 A.M., E.S.T.

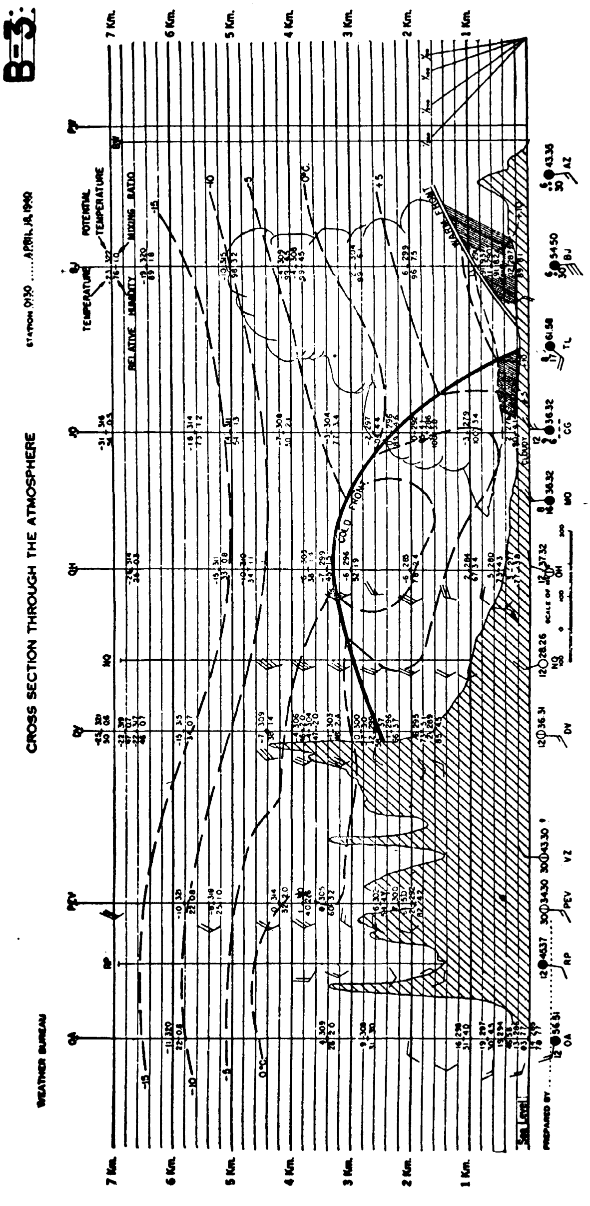


FIG. 130.—Cross section through the atmosphere from Oakland, Calif., to Portland, Me., Apr. 18, 1940, 1.30 A.M., E.S.T.

mainly down the slope of the isentropic surface. This air is therefore relatively dry.

The isentropic charts furnish an invaluable means for the prediction of clouds, thunderstorms, precipitation, icing, etc.; but the methods of analysis and forecasting are somewhat involved and cannot be readily explained without resorting to involved dynamical principles.

CHAPTER XIV

CLIMATE

The word "weather" refers to the more or less instantaneous conditions in the atmosphere or the trend of these conditions over a relatively short period of time. The word "climate" refers to the *mean* or *normal* conditions over a long period, such as 20 or 30 years or more. Thus, although wind, temperature, humidity, cloudiness, etc., are subject to incessant variations, the climate in a given place is more or less invariant.

There is evidence to show that the climate has changed considerably in past ages, and recent observations show that a change is going on now, with a tendency for the winters to become less and less cold. These changes are, however, so slow that the change in climate from one year to the next is negligible. However, if the change in climate continues over long intervals of time, the net result will be considerable. In recent years the word climate has been taken to refer to the mean, or normal, conditions averaged over a 30-year period; and the change in the mean values from one such period to the next expresses the climatic change.

The Elements.—The climate of a place is characterized not only by the long-term annual mean of any single element (such as temperature or rainfall), but by a combination of the normal values of all significant elements. Furthermore, the climate of a place is characterized not only by the mean annual values of the various elements, but also by their normal daily and yearly variations, their extreme values, and the coincidence or noncoincidence of their extremes. Thus, for example, it is important whether the diurnal and annual variations in temperature and rainfall are large or small and whether the rainfall occurs in the warm or in the cold season or whether it is evenly distributed. The most important climatic elements are temperature, precipitation, humidity, cloudiness, etc.; their normal periodic and aperiodic variations; and their extreme values, etc.

Because of the intimate relation that exists between climate and vegetation, climates are to a large extent classified according to the type of plants that grow on uncultivated soil. We may therefore speak of tropical-forest climate, desert climate, pine-forest climate, tundra climate, etc.

The Factors.—The most important external factors that enter into the discussion of climates are

1. *The latitude* of the place in question, for on it depends the angle of incidence of the incoming radiation from the sun, the length of day and night, the length of the seasons, the amount of incoming radiation, etc.

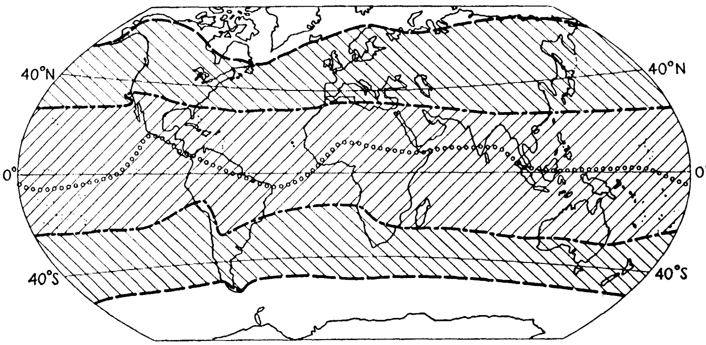


FIG. 131.—Zonal arrangement of certain temperature characteristics.

2. *The elevation* of the place in question, for most meteorological elements vary rapidly with height above sea level.

3. *The properties of the soil*, whether wet, dry, wooded, snow-covered, etc.

4. *The slope* of the surface, for it influences both the amount of precipitation and the temperature conditions.

5. *The continentality*, or the situation of the place in question, with regard to proximity to oceans and exposure to oceanic winds.

Of the above factors, the latitude is the most important, for on it depends the inclination of the sun's rays to the ground. It is perhaps for this reason that the Greek savants chose the word "climate," which literally means "inclination," to denote the mean weather conditions.

This zonal (or latitude) influence is most clearly reflected in the temperature characteristics shown in Fig. 131. It will be seen that the isotherms have a tendency to coincide with the parallel circles with distortions due to the influence of continents and oceans. The so-called *heat equator*[1] is, on the whole, north of the actual equator, because of the fact that the Northern Hemisphere is more continental than the Southern. A more detailed map would show irregularities due to the elevation and the slope of the terrain.

Diurnal and Annual Variation in Incoming Radiation.—From the point of view of incoming radiation, the earth's surface may be divided into five zones, as shown in Fig. 132, *viz:*

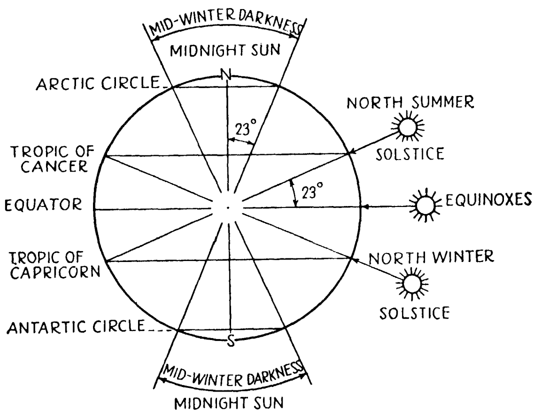


FIG. 132.—Showing the annual variation in the sun's altitude.

1. *The Equatorial Belt.*—The sun is north of the equator during the northern summer and south of the equator during the southern summer. At the equator the sun is in zenith at both equinoxes. About 23° north and south, the sun reaches zenith only at the time of the solstices. Thus, near the equator, the sun is in zenith twice a year, and there will be a maximum of incoming radiation in spring and autumn (see Fig. 133), with the result that there will be a tendency for a double maximum and a double minimum in the annual temperature curve. However, the length of the day varies but little throughout the year, and the sun is high in the sky every day. The annual variation in

[1] The line along which the highest mean annual temperature is observed.

temperature is therefore very small. On the other hand, the diurnal variation in temperature will be relatively large because the length of the day varies but little and the sun rises high in the sky every day.

2. *The Temperate Zones.*—Here, the sun does not reach the zenith in midsummer. The days are long and the sun is high in the sky in summer, and the days are short and the sun is low in winter, with the result that the incoming radiation varies considerably through the year (see Fig. 133). As a result, the annual variation in temperature tends to increase from the equator toward the poles. Simultaneously, the diurnal variation in temperature tends to decrease with increasing latitude, because the midday altitude of the sun decreases.

3. *The Polar Zones.*—On the polar sides of the polar circles, the sun is below the horizon day and night in midwinter and above the horizon day and night in midsummer. At the poles, there is no diurnal variation in the incoming radiation, and the daily variation in temperature vanishes. On the other hand, the difference between the incoming radiation in winter and summer has increased to a maximum (see Fig. 133), with the result that the annual variation in temperature increases.

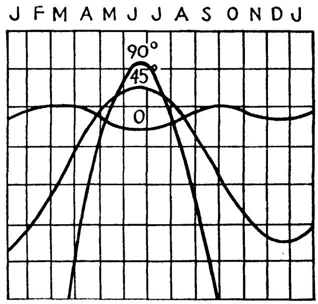


FIG. 133.—Showing the annual variation in incoming radiation at the top of the atmosphere at the equator (0°), 45°N, and at the north pole (90°).

It will be seen from Fig. 133 that the incoming radiation at the top of the atmosphere in summer is greater at the poles than at the equator. This is due to the fact that the sun is above the horizon day and night at the pole, whereas, at the equator, day and night are of about equal length. However, the amount of incoming radiation that reaches the earth's surface depends on the length of the path through the atmosphere. In high latitudes the rays have a long path through the atmosphere, with the result that the incoming radiation at the earth's surface is weakened. Table XI shows how the path of the rays increases as the altitude of the sun above the horizon decreases.

TABLE XI.—LENGTH OF PATH OF INCOMING RADIATION THROUGH THE ATMOSPHERE

Altitude of Sun, Deg.	Length of Path
90	1.00
60	1.15
30	2.00
10	5.70
0	44.70

If the earth's surface were uniform so that the incoming radiation would be absorbed equally and if there were no winds to transport heat from one place to another, the temperature distribution would be controlled by radiation only. The temperature distribution near the earth's surface would then be as follows:

1. *Mean annual temperature* highest at the equator and lowest at the poles.

2. *Annual variation* in temperature small at the equator (with maximum temperature in spring and autumn) and increasing with increasing latitude (with a maximum of temperature in summer).

3. *Diurnal variation* in temperature greatest at the equator and decreasing with increasing latitude.

These rules are approximately true, but they are modified to a considerable extent, mainly for two reasons: (*a*) the distribution of land and sea, and (*b*) the wind systems that transport heat from one place to another.

The Influence of Oceans and Continents on the Air Temperature.—It was explained in a foregoing chapter (page 76) that the influence of oceans on the air temperature tends to reduce the periodic variations. Thus, during the summer (and during the day) the oceans accumulate heat, and during the winter (and during the night) they give off heat to the atmosphere, thus causing the air temperature to vary but little. This may be expressed in the following rule:

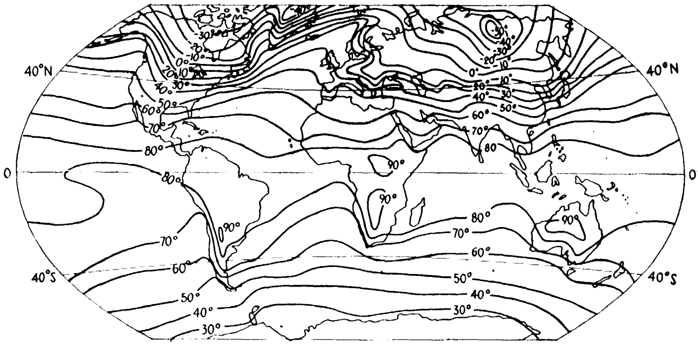
1. The diurnal and annual variations in temperature are small over oceans and increase with increasing distance from the coasts.

How far inland the oceanic influence makes itself felt depends on the prevailing winds. A mere inspection of the wind maps (Figs. 73 and 75) will convince the reader that the following rules hold:

2. In the region of the prevailing westerlies the oceanic influence is greater in the western part of the continents than in the eastern part.

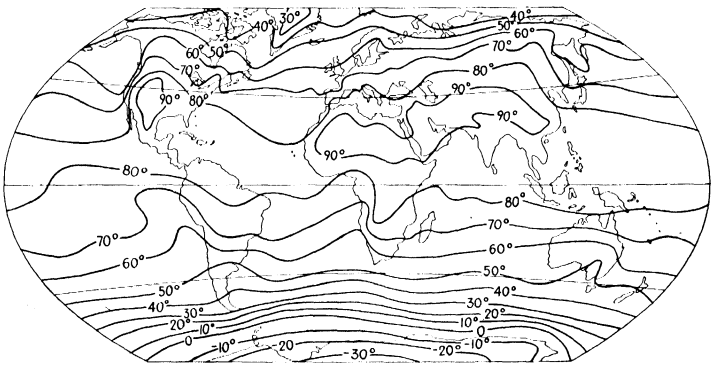
3. The mean annual temperature will, in general, be higher along west coasts than along east coasts in middle latitudes.

4. In the trade-wind regions, the winds are predominantly from the east, and the diurnal and annual variations are larger along west coasts than along east coasts.



MEAN ISOTHERMS-JANUARY (°F)

FIG. 134.



MEAN ISOTHERMS JULY (°F)

FIG. 135.

5. In the monsoon regions (*e.g.*, south Asia) the cold off-land winds in winter lower the temperature, with the result that the annual variation of temperature is above the normal for the latitude.

From the foregoing discussion it follows that temperature conditions near the earth's surface are controlled by a zonal system requiring constant temperature along the parallel circles and by the influences caused by the prevailing winds and the distribution of land and sea. In the free atmosphere, the influence of land and sea tends to vanish with increasing elevation, and the isotherms are more parallel to the latitude circles than they are at the ground.

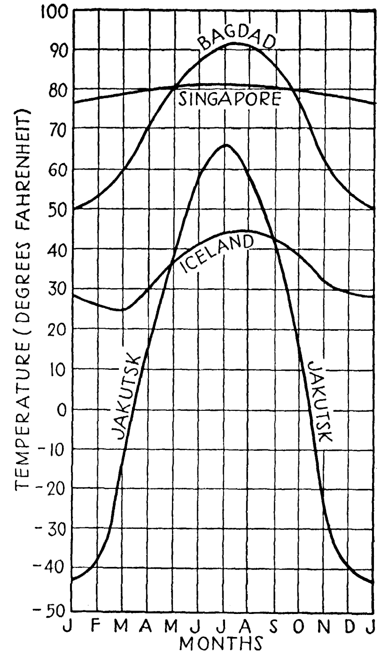


FIG. 136.—Annual variation in temperature. Note how the annual variation varies with increasing latitude from Singapore (lat. 1°N) to Bagdad (lat. 33°N) to Yakutsk (62°N). Note also the difference between Iceland (oceanic) and Yakutsk (continental).

Figures 134 and 135 show the mean air temperature near the earth's surface in the Northern Hemisphere in winter and summer, respectively. Examples of the annual variations in temperature for various latitudes and degrees of continentality are seen in Figs. 136 and 137.

Normal Distribution of Rainfall.—In discussing the distribution of rainfall, the following principles should be borne in mind:

1. The oceans are the main moisture supply for the atmosphere. Through evaporation of water from the oceans and through the condensation and precipitation processes, water is transported from the oceans to the interior of the continents.

2. The amount of moisture that the air can hold depends on the temperature, as shown in Fig. 9. Therefore, during the winter, when the temperature in the interior of the continents is extremely low, the amount of moisture in the atmosphere is so low that no appreciable precipitation can result. In summer, the temperature is high over the northern continents and the air is frequently unstable. The interior of northern continents will therefore have more precipitation in summer than in winter.

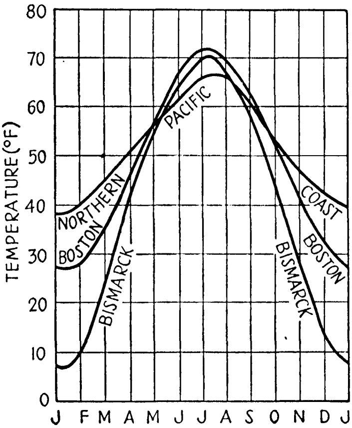


Fig. 137.—In the prevailing westerlies the annual variation in temperature is smaller on a west coast than on an east coast (Boston, Mass.); it is largest inland (Bismarck, N.D.).

3. By far the most effective cause of condensation and precipitation is the adiabatic cooling of ascending air masses. Such ascending motion results in the belts of convergence in the horizontal flow (see Fig. 69) and also on the windward side of mountain ranges.

The principal factors controlling the distribution of precipitation over the earth's surface are, therefore, the belts of convergent flow, the moisture-bearing winds, the air temperature, the distance from the coast, and mountain ranges. Before we discuss the distribution of rainfall in detail, we shall consider the zonal distribution that would prevail if the earth were uniform.

Figure 138 shows a meridional cross section through the atmosphere from the North to the South Pole (see also Figs. 2, 70, and 90). The main zones of convergent flow are the doldrums and the polar-front belt. Here the air ascends, and precipitation occurs in considerable amounts. The main regions of divergent air currents are the subtropical anticyclones. Here the air descends, heats up adiabatically, and becomes relatively dry.

These systems of circulation have an annual rhythm, moving north and south toward the summer hemisphere. Figure 138*A* shows diagrammatically the conditions in summer, and Fig. 138*B* the conditions in winter. Figure 138*C* shows the zones of precipitation that would result from this annual migration of the systems of circulation.

Zone 1.—In the doldrums the temperature is high and the winds are convergent. This results in excessive precipitation.

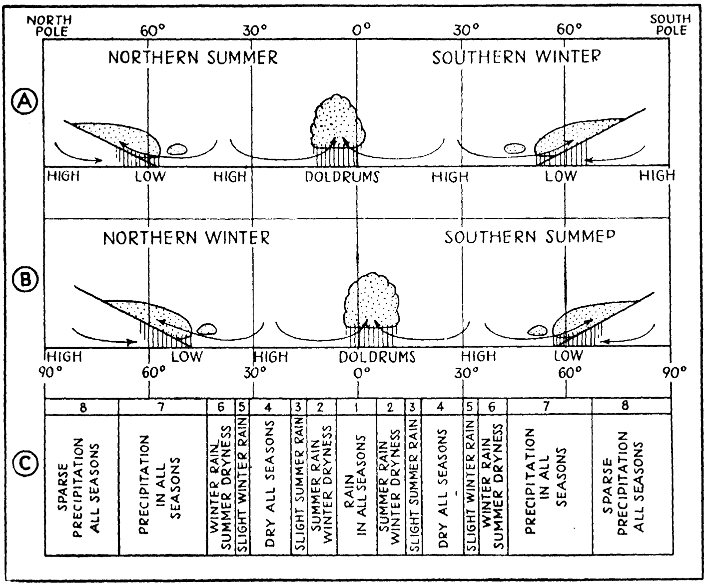


FIG. 138.—Schematical cross section through the atmosphere showing the main zones of ascending and descending motion; *A*, during the northern summer; *B*, during the northern winter; *C*, the zones of precipitation.

In regions where the doldrums are more or less stationary, there will be heavy rainfall in all seasons. However, where the migration is considerable, the rain belt may pass twice a year, causing a double maximum of rainfall to occur. There will therefore be a zone near the doldrums with *heavy rainfall most of the year.*

Zone 2.—In zone 2 in Fig. 138*C* the doldrums will be present in midsummer when the doldrums are farthest away from the equator. In the Northern Hemisphere the wettest season will be

June to August and in the Southern Hemisphere December to February. On the equatorial side, the limit between zones 1 and 2 is drawn where the winter dryness noticeably affects the vegetation. Zone 2, therefore, is characterized by *equatorial summer rain and winter dryness.*

Zone 3.—Regions farther away from the equator, which are visited by the rain from the doldrums only to a slight extent, have *slight rainfall in summer with pronounced dryness the rest of the year.*

Zone 4.—These zones lie under the subtropical anticyclones. Here the descending air currents cause dry weather to prevail. These anticyclones move northward and southward as do the doldrums. In summer, when the anticyclones are farthest away from the equator, the dry region is displaced northward and hence the rainfall in zone 3. Conversely, in winter when the doldrums and the subtropical anticyclones move toward the equator, the prevailing westerlies migrate in the same direction. However, the prevailing westerlies do not reach into zone 4, which therefore remains *dry throughout the year.*

Zone 5.—In midwinter, when the prevailing westerlies spread to lower latitudes, slight amounts of rainfall will occur in zone 5. This zone is therefore characterized by *slight winter rain with pronounced dryness the rest of the year.*

Zone 6.—Farther away from the equator, the duration of the winter rain will be longer than in zone 5. However, in summer this zone is covered by the subtropical anticyclones which then are farthest away from the equator. This zone, which is found to the north of the mean position of the subtropical anticyclones, is characterized by *winter rain and summer dryness* (for example, Southern California).

Zone 7.—This zone is occupied most of the year by the prevailing westerlies and is frequently swept by the polar front. As a result, *rain occurs in all seasons.*

Zone 8.—This zone is occupied by the polar regions of ice and snow; it is frequently visited by fronts and cyclones, but because of the low temperatures the amount of precipitation is moderate or slight.

With this zonal arrangement of the rainfall belts in mind, we now turn to Fig. 139 which shows the mean annual rainfall in both hemispheres. Looking first at Africa, Europe, and Asia, we see that there is a zone of maximum rainfall in the doldrum region. This zone migrates north and south, giving rise to the annual rhythm described above. A zone of pronounced dryness extends from the west coast of north Africa through Arabia into central Asia. Farther north, we meet with the rain belt of the prevailing westerlies. Since these migrate southward in winter, the Mediterranean region obtains most of its rainfall during the winter months. Farther north, rain occurs at all seasons, but the amount decreases as we go inland. The heavy rainfall in south and east Asia is mainly due to the summer monsoon (see page 111).

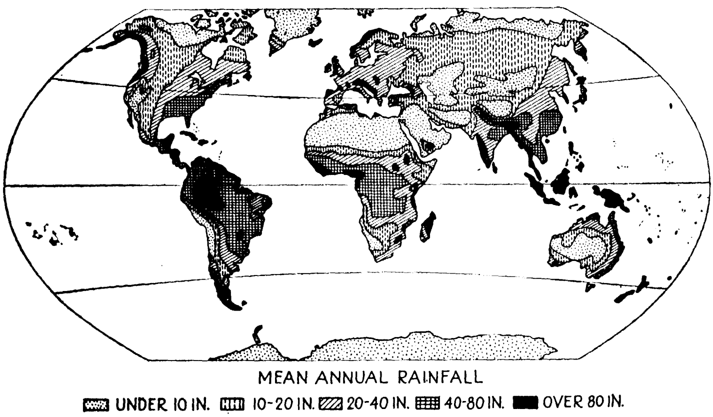


Fig. 139.

Except for the coastal regions, east Asia has relatively dry weather during the winter months.

In the Western Hemisphere, we meet with a similar arrangement, particularly along the west coasts of North and South America. Figure 140 shows clearly the migration of the rain belts along west coasts. In summer, equatorial rain extends into Mexico, and the rain belt of the prevailing westerlies is far to the north. In winter the southern rain belt recedes, and the northern belt has moved down as far as Southern California. Along the east coast of North America, the prevailing winds are on land during the summer (see Fig. 75), and in winter the polar front (see Fig. 88) is normally so close to the coast that precipitation occurs frequently. The subtropical dry belt is therefore not present near the east coast. It is worthy of note that in the region of the prevailing westerlies, the west coasts have more rainfall than the east coasts.

The rainfall distribution in the Southern Hemisphere shows the same zonal arrangement, with dry west coasts and wet east coasts

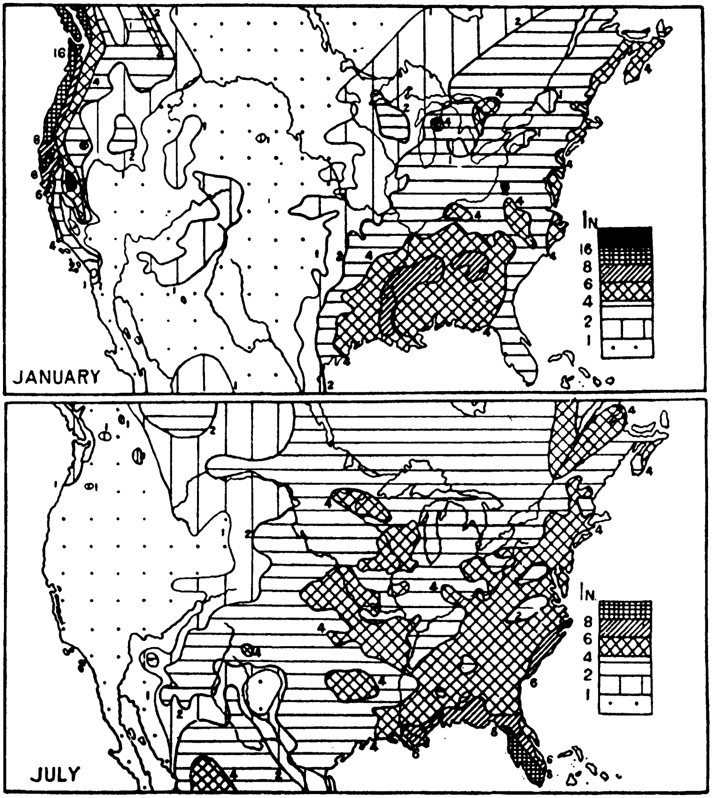


FIG. 140.—Mean monthly rainfall in North America in July and January. Note the migration of the rain belts along the west coast.

in subtropical latitudes and wet west coasts and relatively dry east coasts in high latitudes. It will also be seen from Fig. 139 that the annual amount of rainfall is excessive on the windward slopes of mountain ranges.

Classification of Climates.—A classification has been given by Köppen, who distinguishes among 11 principal types of climate with several subdivisions. We shall here discuss only the 11 principal zones.

In order to understand the principles that govern the geographical distribution of the various climatic zones, we shall consider an idealized continent, as shown in Fig. 141. It will be seen that this continent, in principle, resembles the land masses

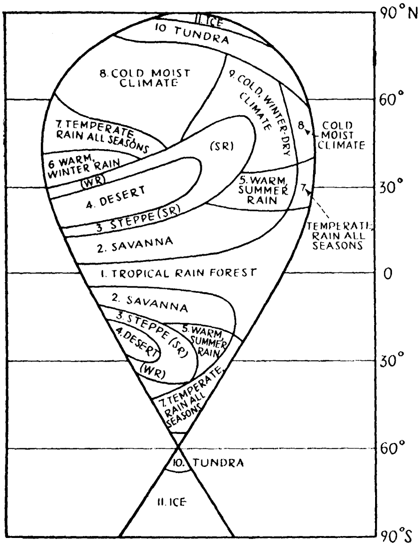


FIG. 141.—Showing the distribution of the principal climatic zones on an idealized continent. (*After Köppen.*) *SR* means summer rainfall; *WR* means winter rainfall.

of Europe, Asia and Africa, with the antarctic continent in the lower portion of the diagram. Naturally, there is no distinct border between the various climatic zones; but the limits between them may be drawn where there are important changes in the natural vegetation.

1. *Tropical-rain-forest Climate.*—This zone occupies the major portion of the equatorial belt (doldrums). Along the west coast the belt is relatively narrow on account of the convergence of the trade winds. Along the east coast, it spreads out to approximately 26°N. and S. because of the monsoons and the on-land

trade winds which give warm weather and rainfall most of the year. This type of climate is characterized by

a. High temperature, coldest month above 18°C. (64.4°F.), annual variation in temperature less than 6°C. (11°F.).

b. Sufficient rainfall to maintain tropical forest, either rain at all seasons, two rain maxima, or one long rain period and one short and dry season[1] with at least 6 cm. (2.4 in.) rainfall in the driest month.

c. Vegetation of the megatherm type, which requires high and constant temperature, abundant precipitation, and high relative humidity.

2. *Tropical-savanna Climate.*—This zone surrounds the tropical rain forests. These regions have a dry period caused by the migration of the doldrums, and the climate is characterized by

a. High temperature, coldest month above 18°C. (64.4°F.), annual variation in temperature less than 12°C. (21.6°F.).

b. Relatively abundant rainfall in summer, and relatively dry winter, with at least 1 month with less than 6 cm. (2.4 in.) rainfall.

c. Vegetation related to the tropical-rain-forest type, but because of the winter dryness the forests are replaced by open land with trees.

3. *Steppes.*—On their poleward sides the savannas merge gradually into semiarid regions where, because of the migrations of the doldrums and the subtropical anticyclones, the summer has some precipitation but the winter is distinctly dry. The steppes continue far into the interior of the continent (see Fig. 141) where the dryness is, in part, due to the large distance from the coasts and lack of moisture-bearing winds. The equatorial part of the steppe region has slight summer rainfall (from the doldrums); the eastern part has slight summer rainfall, chiefly because of summer showers; and the portion indicated by *WR* has dry summer and slight winter rainfall because of the southward migration of the prevailing westerlies in winter. The steppe climate is characterized by

a. Temperature varying within wide limits.

b. Lack of rainfall, evaporation exceeding precipitation, most rain falling as showers at rare intervals, and the amount of rain-

[1] The seasonal variation in rainfall is due to the migration of the doldrums.

fall per year varying considerably, depending on the frequency of precipitation and the season within which it occurs.

c. Vegetation adapted to high temperature, large temperature variations, and long dry periods.

4. *Deserts.*—The desert regions are found in the subtropical latitudes in the western part of the continents where the descending air in the subtropical anticyclones causes extreme dryness. The subtropical deserts do not traverse the continent toward the east coast, for the trade winds and the monsoons along the east coast are moisture-bearing winds. The deserts are characterized by

a. High summer temperatures, large diurnal variation, and moderate annual variation in temperature.

b. Cloudless or almost cloudless sky, extreme dryness, dust and sand storms, rain squalls at very rare intervals.

c. Very sparse vegetation of the steppe type.

5. *Warm Climate with Dry Winter.*—These climatic zones are adjacent to the savannas and occupy regions in low and middle latitudes where the temperature is lower than that of the savannas. The winds are mainly of the monsoon type; the winter is therefore dry and the summer wet. This climate is characterized by

a. Mean temperature of the coldest month below 18°C. (64.4°F.) but above −3°C. (26.6°F.); mean temperature of warmest month over 10°C. (50°F.). (The equatorward limit of snow cover sufficient to protect vegetation during the winter coincides roughly with the −3°C. isotherm in the coldest month.)

b. Dry winter and wet summer; at least ten times as much rainfall in the wettest month of summer as in the driest month of winter.

c. Vegetation of the mesothermal type, adapted to moist and warm summers and dry and moderately cold winters without snow cover.

6. *Warm Climate with Dry Summer.*—These zones lie under the poleward part of the subtropical anticyclones where, because of the annual migration of these anticyclones, the prevailing westerlies give rain in winter. This type of climate, which is often referred to as the Mediterranean type, is characterized by

a. Temperature conditions as in climatic zone 5.

b. Dry summer and moist winter, with at least three times as much rainfall in the wettest month of winter as in the driest month of summer, and the driest month of summer having less than 3 cm. (1.2 in.) of rainfall.

c. Vegetation of the mesothermal type adapted to dry and warm summers and moderately cold and wet winters. The summer is frequently too dry and the winter too cold for the vegetation. As a result, most plants blossom in spring or autumn when there are sufficient heat and moisture.

7. *Humid Temperate Climate.*—These zones (see Fig. 141) are essentially under maritime influence throughout the year, with a moderately high temperature in winter and sufficient rainfall in all seasons. They are characterized by

a. Temperature as in climatic zones 5 and 6.

b. No appreciable annual variation in rainfall.

c. Vegetation of the mesothermal type adapted for high moisture throughout the year (evergreens).

8. *Cold Climate with Moist Winter.*—These regions coincide with the subpolar belts of pine forest. On the western side of the continent in Fig. 141 a vast area is occupied by this zone, whereas only a relatively small area is present along the east coast. This is due to the fact that the prevailing westerlies bring moisture and precipitation far inland in the western part of the continent in winter, whereas the moist area along east coasts in high latitudes is limited to the coastal regions. Climatic zones 8 are characterized by

a. Mean temperature of coldest month less than −3°C. (26.6°F.), and mean temperature of warmest month above 10°C. (50°F.).

b. Precipitation throughout the year; on the coasts mostly in winter, and inland mostly in summer, without any season being particularly dry.

c. Vegetation principally in the microthermal type which requires relatively short summers and long winter rest and needs snow cover for protection during the long and cold winters (*e.g.*, pine and fir.)

9. *Cold Climate with Dry Winter.*—This zone occupies the most continental part of the continent in high latitudes. Because of the low winter temperature and the great distance from moisture-bearing winds, the amount of precipitation during the winter months is exceedingly small. Otherwise, the characteristics are similar to those of zone 8.

10. *Tundra Climate.*—This zone occupies the northernmost part of the continent. The mean temperature of the warmest month is below 10°C. (50°F.). The subsoil is frozen throughout the year, and there are no forests.

11. *Ice Climate.*—The polar cap of permanent snow and ice, with mean temperature of the warmest month below 0°C. (32°F.).

It will be seen that the distribution of the climatic zones on the idealized continent (Fig. 141) results logically from the superimposition of the rainfall zones (see Figs. 138 and 139) on the temperature distribution (Figs. 134 and 135). In low latitudes, the temperature is sufficiently high to maintain vegetation; here the rainfall is the main control. In middle latitudes, temperature and precipitation are of equal importance for the maintenance of vegetation, and in high latitudes the vegetation is mainly controlled by the temperature.

If we travel along the west coast from the equator toward the North Pole, we pass the climatic zones in the following order: tropical forest (1), savanna (2), steppe (3) with summer rain, desert (4), steppe (5) with winter rain, warm climate with wet winter (6), warm climate with rain in all seasons (7), cold climate with moist winter (8), tundra (10), and ice (11).

If we travel along the east coast from the equator toward the North Pole, we pass the climatic zones in the following order: a wide zone of tropical forest (1), merging gradually into warm climate with rain in all seasons (7), cold climate with moist winter (8), tundra (10), and ice (11).

South of the equator the distribution of the climate is principally the same, with the exception that the cold climates 8 and 9 are missing because of the absence of extensive land areas in subpolar latitudes.

Köppen's classification, as applied to actual observations, is shown in Fig. 142. The borders between the various zones are somewhat distorted because of the varying elevation and the irregular shape of the land areas; but in the main there is excellent agreement between the distribution of the climates on the idealized continent and what is actually observed.

CHAPTER XV

HISTORY

The Background.—The history of meteorology has many features in common with the history of other sciences. Usually a new science commences with a few more or less accidental observations of certain phenomena. The interest thus created usually results in an organized program of observations, followed by a systematic working up of observations, more or less along statistical lines. Eventually, the theorist tries to extract from the observations and the empirical rules the laws of nature that govern the phenomena, and this marks the transition from the descriptive to the exact epoch in the development of the science in question. Thus, in the old times the astronomers commenced to observe the movement of the heavenly bodies. Through the working up of long series of observations, Copernicus and Kepler were able to extract certain empirical rules for the movement of the planets; and, finally, Newton, through his discovery of the law of gravitation, replaced the empirical rules by laws of nature and changed astronomy from a descriptive and statistical science to an exact one.

The development in meteorology has not been so rapid as in the sister science, astronomy. However, in view of the rapid advances in recent years, it seems safe to say that the science of meteorology is now in a state of transition from the descriptive to the exact state, although the road to complete exactitude may be long and winding.

Fundamentally, meteorology is physics, mathematics, mechanics, and chemistry applied to the atmosphere. The development in meteorology, therefore, has depended on the developments in these sciences. Only after the discovery of the principles of mechanics and the invention of such basic instruments as the barometer and the thermometer could real progress be made.

Since it is obviously difficult to isolate the phenomena in the atmosphere or to perform controlled laboratory experiments bearing on large-scale atmospheric processes, it was necessary to study the atmospheric processes by means of simultaneous observations from vast areas. Such observations could be obtained only after the invention of telegraphy. This invention, therefore, marks an epoch in the history of meteorology. A further and considerable advance was made when the radio and teletype came into general usage.

Finally, the funds required for the provision of the vast mass of observational material are so considerable as hardly to be forthcoming unless meteorology could satisfy specific public demands. Although the value of meteorology for many human activities has always been recognized, it was only through the development of aviation that the demands rose to such proportions that reasonably adequate funds became available for synoptic exploration of the atmosphere in all three dimensions.

Although the sum total of expenditures for meteorology may be considerable, it is perhaps not out of place to mention that the expense per capita per year is of the order of magnitude of one or two postage stamps.

From Aristotle to Galileo.—The earliest indications of scientific activity in the field of meteorology go back, at least, to the fifth century B.C. with more or less regular visual observations of certain phenomena. About the year 400 B.C., Hippocrates wrote the first treatise on medical climatology with many interesting inferences. About 350 B.C. Aristotle wrote the first text on meteorology based on his own observations. It is known that rain measurements were made in India as early as in the fourth century B.C. and that wind vanes were used in the first century B.C. The development was not, however, very rapid, for no appreciable progress is noted until the ninth century A.D. when weather vanes were used quite frequently on church steeples.

Perhaps the first instrument with a reacting substance was made by Cardinal de Cusa in the fifteenth century, who determined the humidity by weighing balls of wool under various moisture conditions. About 1500, Leonardo da Vinci constructed an improved wind vane and a mechanical moisture indicator. In 1597, Galileo made a temperature indicator, which later became a measuring instrument through the addition of an arbitrary scale. This marks the transition from visual to instrumental observations, or, in other words, the first step toward exactitude.

From Galileo to Leverrier.—The works of Galileo and his students mark an epoch in the history of physics and mechanics. In 1643, Torricelli made the first mercurial barometer, and five years later Periers, following a suggestion by Pascal, Descartes, and others, made the famous barometer observations on Puy-de-Dôme to show that the atmospheric pressure decreases with height.

From 1650 to 1850, instrumental developments followed in rapid succession. It suffices here to mention the following events: the condensation hygrometer by Ferdinand II of Tuscany (1650), the Florentine thermometer (about 1600), Wren's meteorograph (1664), the introduction by Huygens of the freezing point and the boiling point of water as reference points on the thermometer scale (1665), the Fahrenheit scale (1710), the Reaumur scale (1733), the Celsius (or Centigrade) scale (1736) which originally had 0° at the boiling point and 100° at the freezing point, the lightning conductor by Franklin (1749), the hair hygrometer by De Saussure (1783), the anemometer by Woltman (1790), the Beaufort scale for wind force (1805), the psychrometer by August (1825), the pyrheliometer and the definition of the solar constant by Pouillet (1837), and the aneroid barometer by Vidie (1847).

The development of instruments and the observations made during this period gave rise to many important discoveries. In that era of sailing vessels the winds were of particular importance. In 1624 the rotation of the winds with the sun was discovered by von Verulam. In 1686, Halley gave the first account of the trade winds and the monsoons. After 1846, through the use of the maps of the prevailing winds and ocean currents prepared by Maury, the sailing time from England to Australia and return was reduced from about 250 to about 150 days.

Attention was soon directed to such traveling disturbances as cyclones, and tropical storms. Already in 1687, Dampier recognized that the typhoons were revolving storms, and about 1840 Redfield found the same to be true of other cyclones. About 1840, Dove published his theory of storms, with indications that storms originate when polar air and equatorial (tropical) air are brought into juxtaposition. This was followed by Ferrel's theory of the general circulation (1856).

Of great theoretical importance were Dalton's discovery of the gas law (1793), Carnot's studies of heat (1826), and various studies in radiation. But, on the whole, the period from 1650 to 1850 is characterized by statistical investigations resulting at the end of the period in primitive maps of the mean distribution of some of the meteorological elements over the earth's surface.

The earliest indications of synoptic weather charts go back to about 1820 (Brandes), but it was only after the Crimean War (1854 to 1856) that weather maps came into regular use through the initiative of Napoleon III. The following period is therefore colored with the experience gained from daily synoptic weather charts and with a slow but steady infiltration into meteorology of the physical laws that govern motion and the relation between heat and motion. Thus, through Watt's invention of the steam engine (1765) and Carnot's book "Réflexions sur la puissance motrice du feu" (1826), the era of the so-called "motor age" commenced, leading to our present-day aviation with its profound influence on meteorology.

From Leverrier to Bjerknes.—During the Crimean War a strong storm arrived in the Black Sea and caused considerable damage to the fleet, particularly to the French battleship *Henri IV*. About this time, the astronomer Leverrier had won world fame by forecasting the existence of a new planet. From Newton's law of gravitation and from the observed orbits of the known planets, Leverrier computed that another planet must exist within our solar system. He also computed where this planet would be in the sky at a certain time. The telescopes were directed to the computed spot, and, behold, the planet was found and named Neptune.

The French emperor Napoleon III apparently thought that, if science could predict the whereabouts of hitherto unknown planets, it could also forecast storms and weather. He therefore charged Leverrier with the, as it proved, difficult task of organizing a system of weather forecasting.

At this time, there were no Weather Bureau offices to turn to for information, no organized network of stations and system of reporting. But from the observatories, universities, and the few observing stations keeping meteorological logs, Leverrier collected so many data that he could plot primitive weather maps by means of which he could study the Black Sea storm "post mortem." He then found that the storm could be traced from one weather map to another as the storm developed and moved on a regular path and with a fairly constant speed toward the Black Sea. The conclusion drawn was that, if observations were made at a large number of reporting stations and if the reports could be transmitted with sufficient speed to a central office, one could, by plotting and analyzing weather charts, follow the storm from chart to chart and extrapolate its future movement.

By this, Leverrier had not solved any problem related to the physics or the dynamics of the storms; he had merely shown that a simple freehand extrapolation of the storm path could render valuable and practical results.

These results created a wave of enthusiasm. Within a few years, meteorological institutes and a network of stations were established in most countries, and the experiments with storm warnings and weather forecasting commenced. However, it was soon found that the problem of forecasting was more intricate than anticipated. The storm that Leverrier had investigated was a pronounced and clear-cut case. But it happened frequently that the indications were confusing; the atmospheric phenomena seemed to develop around an unstable state of equilibrium so that a small impulse could give rise to great events; and it was difficult to trace the phenomena back to their ultimate, or penultimate, causes. Furthermore, at this time, little was known of the mean, or normal, state of the atmosphere; and before the normal state was known, it was difficult to understand the perturbations superimposed thereon.

The methods of forecasting that developed were mostly based on statistical rules, rules of thumb, etc. These rules were mostly of local or regional value, with the result that a forecaster in London could learn but little from his colleague in Washington, and vice versa.

The optimism created by Leverrier's report was soon followed by pessimism, and the problem of weather forecasting on a rational scientific basis was not seriously attacked until about the time of World War I.

The lack of development within the field of weather forecasting was, however, offset by an intense activity in descriptive and statistical meteorology, and many of the empirical results won during this period formed the basis for later theoretical researches.

In 1860, Buys-Ballot discovered the so-called baric wind law, which states a relation between the wind and the pressure distribution. The first map of the pressure distribution over the entire world was produced by Buchan in 1869, followed by a storm atlas by Mohn in 1870. The storm tracks of cyclones were mapped by Jackson, Köppen, von Bebber and others (1878 to 1882). The first world map of precipitation was prepared by Loomis in 1882; the periodic and aperiodic variations of the various meteorological elements were studied by numerous authors, and in 1887 the first meteorological atlas appeared (Hann); the ideas of the normal conditions of the atmosphere near the earth's surface and the distribution of the climatic zones began to take definite form.

The first scientific balloon ascent was made in Paris by Charles as early as 1803, and several ascents were made by Glashier from 1862 to 1866. From 1890, several mountain observatories were erected to record the conditions in the free atmosphere, and balloons and kites carrying instruments were used with increasing frequency to obtain observations from the upper atmosphere. In 1901, Suring and Berson reached the altitude of 10,800 m. (35,400 ft.) in a free balloon, a year later Teisserenc de Bort and Assmann discovered the stratosphere, and Gold and Humphreys soon gave a theoretical explanation for the formation of the tropopause. However, simultaneous observations from the free atmosphere in large numbers and over large areas were obtainable only after the airplane and the radio came into general use.

The vast uninhabited regions in the arctic and antarctic were tempting fields for exploration, and many valuable data were brought home by the numerous polar expeditions (notably by Nansen, Scott, Amundsen, and Sverdrup) to enhance our knowledge of the meteorology of these regions.

During the period preceding World War I, a considerable amount of theoretical investigations of fundamental importance were made. Thus, theoretical contributions to our knowledge of atmospheric motion were made by Buys-Ballot (1860), Guldberg and Mohn (1877), Sprung (1885), Helmholtz (1888), Ferrel (1889), Margules (1803), Ekman (1905), V. Bjerknes (1898 and 1912), Hesselberg and Sverdrup (1914), and Exner (1917). Simultaneously, important contributions to the thermodynamics of the atmosphere were rendered by Hertz (1884), Helmholtz (1889), Bezold (1889), Emden (1907), and others.

Considering this period in retrospect, it appears that not only were the theorists in the minority, but little attention was paid to their results. This is particularly true of Helmholtz's papers on atmospheric motion which contain many interesting inferences that have not yet been fully evaluated. The same is true also of Bjerknes's circulation theorem, published in 1898. Although this theorem greatly accelerated the development in dynamic oceanography, its influence on dynamic meteorology is of more recent date.

Another feature of considerable interest is the early ideas of wave motion in the atmosphere. The wave-generating forces investigated originally were the gravity force and the wind shear (Helmholtz's waves), and it was found that waves could be stable or unstable. However, the inclusion of the inertia force due to the earth's rotation in the wave theory is of recent date (Solberg, (1928) and has led to a more complete understanding of the development of cyclones. But this brings us to the most recent and fruitful period in the history of meteorology.

From World War I to World War II.—This period is so recent that it is too early to record its history. Perhaps the most outstanding achievements during the last 25 years are the discovery of the polar front, the wave theory of cyclones, the air-mass and frontal methods of weather forecasting as initiated by V. Bjerknes and his collaborators in Norway, and the method of isentropic analysis as initiated by Rossby and his collaborators in the United States.

In the other branches of meteorology, the period under discussion has been equally fruitful. Thus, the theories of turbulence and heat transfer have been greatly developed. The same is true also of the theories of convection, radiation, condensation and precipitation processes, etc. A feature of singular interest is the rapid advances made in the observation technique. The airplane meteorograph, and even more so, the radiometeorograph have made it possible to obtain simultaneous observations from the free atmosphere that have greatly enhanced our knowledge of the atmospheric processes and increased the accuracy of weather forecasting.

With our present methods, weather phenomena may be forecast with reasonable accuracy for 24 to 36 hr. in advance. The problem of extending the forecasts to a week, a month, or a season in advance has been attacked by many authors. Thus, in India, Sir Gilbert Walker has developed a correlation method for forecasting the variations in the monsoon rainfall. In Germany, Baur makes forecasts for 5 to 10 days in advance, and in the United States Rossby and his collaborators are developing a method for weekly forecasts that has yielded highly promising results.

In view of the rapid development during the last 25 years, it seems reasonable to expect that the accuracy of our short-term forecasts will continue to improve and that methods of long-range forecasting will be developed within the near future. Weather forecasting, which was actually initiated through the events that occurred during the Crimean War and greatly enhanced through the demands that arose during World War I, will probably make further important advances as a result of World War II.

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APPENDIX

CONVERSION TABLES

TABLE I.—CONVERSION OF INCHES OF MERCURY TO MILLIBARS

In.	.00	.01	.02	.03	.04	.05	.06	.07	.08	.09
0.00	0.00	0.34	0.68	1.02	1.35	1.69	2.03	2.37	2.71	3.05
0.10	3.39	3.73	4.06	4.40	4.74	5.08	5.42	5.76	6.10	6.43
0.20	6.77	7.11	7.45	7.79	8.13	8.47	8.80	9.14	9.48	9.82
0.30	10.16	10.50	10.84	11.18	11.51	11.85	12.19	12.53	12.87	13.21
0.40	13.55	13.88	14.22	14.56	14.90	15.24	15.58	15.92	16.25	16.59
0.50	16.93	17.27	17.61	17.95	18.29	18.63	18.96	19.30	19.64	19.98
0.60	20.32	20.66	21.00	21.33	21.67	22.01	22.35	22.69	23.03	23.37
0.70	23.70	24.04	24.38	24.72	25.06	25.40	25.74	26.08	26.41	26.75
0.80	27.09	27.43	27.77	28.11	28.45	28.78	29.12	29.46	29.80	30.14
0.90	30.48	30.82	31.15	31.49	31.83	32.17	32.51	32.85	33.19	33.53
1.00	33.86	34.20	34.54	34.88	35.22	35.56	35.90	36.23	36.57	36.91
1.10	37.25	37.59	37.93	38.27	38.60	38.94	39.28	39.62	39.96	40.30
1.20	40.64	40.98	41.31	41.65	41.99	42.33	42.67	43.01	43.35	43.68
1.30	44.02	44.36	44.70	45.04	45.38	45.72	46.05	46.39	46.73	47.07
1.40	47.41	47.75	48.09	48.43	48.76	49.10	49.44	49.78	50.12	50.46
1.50	50.80	51.13	51.47	51.81	52.15	52.49	52.83	53.17	53.51	53.84
1.60	54.18	54.52	54.86	55.20	55.54	55.88	56.21	56.55	56.89	57.23
1.70	57.57	57.91	58.25	58.58	58.92	59.26	59.60	59.94	60.28	60.62
1.80	60.96	61.29	61.63	61.97	62.31	62.65	62.99	63.33	63.66	64.00
1.90	64.34	64.68	65.02	65.36	65.70	66.03	66.37	66.71	67.05	67.39
2.00	67.73	68.07	68.41	68.74	69.08	69.42	69.76	70.10	70.44	70.78
2.10	71.11	71.45	71.79	72.13	72.47	72.81	73.15	73.48	73.82	74.16
2.20	74.50	74.84	75.18	75.52	75.86	76.19	76.53	76.87	77.21	77.55
2.30	77.89	78.23	78.56	78.90	79.24	79.58	79.92	80.26	80.60	80.93
2.40	81.27	81.81	81.95	82.29	82.63	82.97	83.31	83.64	83.98	84.32
25.00	846.6	846.9	847.3	847.6	848.0	848.3	848.6	849.0	849.3	849.6
25.10	850.0	850.3	850.7	851.0	851.3	851.7	852.0	852.4	852.7	853.0
25.20	853.4	853.7	854.0	854.4	854.7	855.1	855.4	855.7	856.1	856.4
25.30	856.8	857.1	857.4	857.8	858.1	858.5	858.8	859.1	859.5	859.8
25.40	860.1	860.5	860.8	861.2	861.5	861.8	862.2	862.5	862.9	863.2
25.50	863.5	863.9	864.2	864.5	864.9	865.2	865.6	865.9	866.2	866.6
25.60	866.9	867.3	867.6	867.9	868.3	868.6	868.9	869.3	869.6	870.0
25.70	870.3	870.6	871.0	871.3	871.7	872.0	872.3	872.7	873.0	873.4
25.80	873.7	874.0	874.4	874.7	875.0	875.4	875.7	876.1	876.4	876.7
25.90	877.1	877.4	877.8	878.1	878.4	878.8	879.1	879.4	879.8	880.1

TABLE I.—CONVERSION OF INCHES OF MERCURY TO MILLIBARS.—(*Continued*)

In.	.00	.01	.02	.03	.04	.05	.06	.07	.08	.09
26.00	880.5	880.8	881.1	881.5	881.8	882.2	882.5	882.8	883.2	883.5
26.10	883.8	884.2	884.5	884.9	885.2	885.5	885.9	886.2	886.6	886.9
26.20	887.2	887.6	887.9	888.3	888.6	888.9	889.3	889.6	889.9	890.3
26.30	890.6	891.0	891.3	891.6	892.0	892.3	892.7	893.0	893.3	893.7
26.40	894.0	894.3	894.7	895.0	895.4	895.7	896.0	896.4	896.7	897.1
26.50	897.4	897.7	898.1	898.4	898.7	899.1	899.4	899.8	900.1	900.4
26.60	900.8	901.1	901.5	901.8	902.1	902.5	902.8	903.2	903.5	903.8
26.70	904.2	904.5	904.8	905.2	905.5	905.9	906.2	906.5	906.9	907.2
26.80	907.6	907.9	908.2	908.6	908.9	909.2	909.6	909.9	910.3	910.6
26.90	910.9	911.3	911.6	912.0	912.3	912.6	913.0	913.3	913.6	914.0
27.00	914.3	914.7	915.0	915.3	915.7	916.0	916.4	916.7	917.0	917.4
27.10	917.7	918.1	918.4	918.7	919.1	919.4	919.7	920.1	920.4	920.8
27.20	921.1	921.4	921.8	922.1	922.5	922.8	923.1	923.5	923.8	924.1
27.30	924.5	924.8	925.2	925.5	925.8	926.2	926.5	926.9	927.2	927.5
27.40	927.9	928.2	928.5	928.9	929.2	929.6	929.9	930.2	930.6	930.9
27.50	931.3	931.6	931.9	932.3	932.6	933.0	933.3	933.6	934.0	934.3
27.60	934.6	935.0	935.3	935.7	936.0	936.3	936.7	937.0	937.4	937.7
27.70	938.0	938.4	938.7	939.0	939.4	939.7	940.1	940.4	940.7	941.1
27.80	941.4	941.8	942.1	942.4	942.8	943.1	943.4	943.8	944.1	944.5
27.90	944.8	945.1	945.5	945.8	946.2	946.5	946.8	947.2	947.5	947.9
28.00	948.2	948.5	948.9	949.2	949.5	949.9	950.2	950.6	950.9	951.2
28.10	951.6	951.9	952.3	952.6	952.9	953.3	953.6	953.9	954.3	954.6
28.20	955.0	955.3	955.6	956.0	956.3	956.7	957.0	957.3	957.7	958.0
28.30	958.3	958.7	959.0	959.4	959.7	960.0	960.4	960.7	961.1	961.4
28.40	961.7	962.1	962.4	962.8	963.1	963.4	963.8	964.1	964.4	964.8
28.50	965.1	965.5	965.8	966.1	966.5	966.8	967.2	967.5	967.8	968.2
28.60	968.5	968.8	969.2	969.5	969.9	970.2	970.5	970.9	971.2	971.6
28.70	971.9	972.2	972.6	972.9	973.2	973.6	973.9	974.3	974.6	974.9
28.80	975.3	975.6	976.0	976.3	976.6	977.0	977.3	977.7	978.0	978.3
28.90	978.7	979.0	979.3	979.7	980.0	980.4	980.7	981.0	981.4	981.7
29.00	982.1	982.4	982.7	983.1	983.4	983.7	984.1	984.4	984.8	985.1
29.10	985.4	985.8	986.1	986.5	986.8	987.1	987.5	987.8	988.2	988.5
29.20	988.8	989.2	989.5	989.8	990.2	990.5	990.9	991.2	991.5	991.9
29.30	992.2	992.6	992.9	993.2	993.6	993.9	994.2	994.6	994.9	995.3
29.40	995.6	995.9	996.3	996.6	997.0	997.3	997.6	998.0	998.3	998.6
29.50	999.0	999.3	999.7	1000.0	1000.3	1000.7	1001.0	1001.4	1001.7	1002.0
29.60	1002.4	1002.7	1003.1	1003.4	1003.7	1004.1	1004.4	1004.7	1005.1	1005.4
29.70	1005.8	1006.1	1006.4	1006.8	1007.1	1007.5	1007.8	1008.1	1008.5	1008.8
29.80	1009.1	1009.5	1009.8	1010.2	1010.5	1010.8	1011.2	1011.5	1011.9	1012.2
29.90	1012.5	1012.9	1013.2	1013.5	1013.9	1014.2	1014.6	1014.9	1015.2	1015.6
30.00	1015.9	1016.3	1016.6	1016.9	1017.3	1017.6	1018.0	1018.3	1018.6	1019.0
30.10	1019.3	1019.6	1020.0	1020.3	1020.7	1021.0	1021.3	1021.7	1022.0	1022.4
30.20	1022.7	1023.0	1023.4	1023.7	1024.0	1024.4	1024.7	1025.1	1025.4	1025.7
30.30	1026.1	1026.4	1026.8	1027.1	1027.4	1027.8	1028.1	1028.4	1028.8	1029.1
30.40	1029.5	1029.8	1030.1	1030.5	1030.8	1031.2	1031.5	1031.8	1032.2	1032.5
30.50	1032.9	1033.2	1033.5	1033.9	1034.2	1034.5	1034.9	1035.2	1035.6	1035.9
30.60	1036.2	1036.6	1036.9	1037.3	1037.6	1037.9	1038.3	1038.6	1038.9	1039.3
30.70	1039.6	1040.0	1040.3	1040.6	1041.0	1041.3	1041.7	1042.0	1042.3	1042.7
30.80	1043.0	1043.3	1043.7	1044.0	1044.4	1044.7	1045.0	1045.4	1045.7	1046.1
30.90	1046.4	1046.7	1047.1	1047.4	1047.8	1048.1	1048.4	1048.8	1049.1	1049.5

TABLE II.—DECREASE IN PRESSURE (MILLIBARS) PER 100 FT. ASCENT

Pressure, mb.	Temperature, °F.			
	−50	0	50	100
1000	4.6	4.1	3.7	3.4
900	4.1	3.7	3.3	3.0
800	3.7	3.3	3.0	2.7
700	3.2	2.9	2.6	2.3
600	2.7	2.4	2.2	2.0
500	2.3	2.0	1.8	1.7

TABLE III.—CONVERSION FROM CENTIGRADE TO FAHRENHEIT

°C.	0	1	2	3	4	5	6	7	8	9
+40	104.0	105.8	107.6	109.4	111.2	113.0	114.8	116.6	118.4	120.2
+30	86.0	87.8	89.6	91.4	93.2	95.0	96.8	98.6	100.4	102.2
+20	68.0	69.8	71.6	73.4	75.2	77.0	78.8	80.6	82.4	84.2
+10	50.0	51.8	53.6	55.4	57.2	59.0	60.8	62.6	64.4	66.2
+ 0	32.0	33.8	35.6	37.4	39.2	41.0	42.8	44.6	46.4	48.2
− 0	32.0	30.2	28.4	26.6	24.8	23.0	21.2	19.4	17.6	15.8
−10	14.0	12.2	10.4	8.6	6.8	5.0	3.2	1.4	− 0.4	− 2.2
−20	− 4.0	− 5.8	− 7.6	− 9.4	−11.2	−13.0	−14.8	−16.6	−18.4	−20.2
−30	−22.0	−23.8	−25.6	−27.4	−29.2	−31.0	−32.8	−34.6	−36.4	−38.2
−40	−40.0	−41.8	−43.6	−45.4	−47.2	−49.0	−50.8	−52.6	−54.4	−56.2

TABLE IV.—CONVERSION OF METERS PER SECOND TO KILOMETERS PER HOUR TO MILES PER HOUR TO KNOTS

m./sec.	km./hr.	m.p.h.	Knots
1	3.6	2.2	1.9
2	7.2	4.5	3.9
3	10.8	6.7	5.8
4	14.4	8.9	7.8
5	18.0	11.2	9.7
6	21.6	13.4	11.7
7	25.2	15.7	13.6
8	28.8	17.9	15.6
9	32.4	20.1	17.5
10	36.0	22.4	19.4
11	39.6	24.6	21.4
12	43.2	26.8	23.3
13	46.8	29.1	25.3
14	50.4	31.3	27.2
15	54.0	33.6	29.1
16	57.6	35.8	31.1
17	61.2	38.0	33.0
18	64.8	40.3	35.0
19	68.4	42.5	36.9
20	72.0	44.7	38.9
21	75.6	47.0	40.8
22	79.2	49.2	42.7
23	82.8	51.4	44.7
24	86.4	53.7	46.6
25	90.0	55.9	48.6
26	93.6	58.2	50.5
27	97.2	60.4	52.5
28	100.8	62.6	54.4
29	104.4	64.9	56.3
30	108.0	67.1	58.3
31	111.6	69.3	60.2
32	115.2	71.6	62.2
33	118.8	73.8	64.1
34	122.4	76.1	66.0
35	126.0	78.3	68.0
36	129.6	80.5	69.9
37	133.2	82.8	71.9
38	136.8	85.0	73.8
39	140.4	87.2	75.8
40	144.0	89.5	77.7
41	147.6	91.7	79.6
42	151.2	94.0	81.6
43	154.8	96.2	83.5
44	158.4	98.4	85.5
45	162.0	100.7	87.4
46	165.6	102.9	89.4
47	169.2	105.1	91.3
48	172.8	107.4	93.2
49	176.4	109.6	95.2
50	180.0	111.8	97.1

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(Does not include items mentioned in the chapter on History)

D

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K

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M