MANUAL OF METEOROLOGY

PART 1

DEPARTMENT OF SCIENCE
AND TECHNOLOGY
BUREAU OF METEOROLOGY

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PART 1

FOREWORD

The *Manual of Meteorology* has been revised. This part, which is Part 1, dealing with general meteorology, has been prepared with two main purposes in mind:

(a) to form the common basis of a series for those whose work requires an understanding of meteorology, such as aviators, air traffic controllers, fire control officers, and mariners;

(b) to provide a reference book on general meteorology for secondary students and others interested in the subject.

The former *Aviation Supplement* has been revised and issued as Part 2. Further parts will be prepared as required and as resources permit.

A working knowledge of mathematice and physics will be of advantage to the reader of the *Manual of Meteorology*. However, complex mathematical treatment has been avoided in the presentation.

Grateful acknowledgement is made to the World Meteorological Organization for permission to use material from the *Compendium of Lecture Notes for Training Class IV Meteorological Personnel.*

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CONTENTS

CHAPTER 1

THE COMPOSITION OF THE ATMOSPHERE

Most men spend their lives at the bottom of the ocean of air that forms part of our planet. The various gases which surround the earth are bound to it by the gravitational force of attraction.

The upper limit to the atmosphere is sometimes conventionally described as occurring at about 1000 km above mean sea level. However, most scientists prefer to regard it as merging gradually with the extremely rarefied gases and dust that lie between the planets. In this case there is no sharp boundary between the atmosphere and outer space.

Mere consideration of the vertical extent of the atmosphere could, however, be misleading. Even at the conventional upper limit of 1000 km, the atmosphere is so tenuous that it has a density less than that of any man-made vacuum. In fact only about 1% of the mass of the atmosphere lies above an altitude of 30 km.

In order to understand the physical processes that take place in the atmosphere we must first study its composition. This chapter deals with the constituents of the atmosphere and their distribution in space.

1.1 The composition of dry air

The atmosphere consists essentially of a mixture of gases. However, within its gases are suspended tiny solid particles of dust and smoke. In addition, water occurs not only as a vapour, but also in the solid and liquid states.

The composition of dry air by volume at sea level is shown in Table 1.1. This refers to locations remote from large cities and forest fires.

Table 1.1 The composition of dry air

Gas	Percentage by volume
Nitrogen	78.084
Oxygen	20.946
Argon	0.934
Carbon dioxide	0.033
Neon	0.00182
Helium	0.00052
Krypton, hydrogen, xenon, ozone, radon, etc.	0.00066

In general, the gases of the atmosphere are present in the same proportions up to an altitude of about 80 km. There are, however, two important exceptions—ozone (O_3) and water vapour (H_2O). Some variability of carbon dioxide (CO_2) also occurs at low levels.

1.2 Atmospheric ozone

Ozone molecules (O_3) consist of three atoms of oxygen. The concentration of ozone varies with altitude, latitude and time.

Most ozone forms in the upper stratosphere as a result of processes which include the absorption of ultraviolet radiation. The ozone molecules then tend to sink in the atmosphere and accumulate in the lower stratosphere at altitudes between 15 and 25 km.

Small amounts of ozone are also produced in the air near the earth's surface, as a result of electrical discharges. However, the concentration of ozone at any particular altitude varies considerably, due to the circulation of the atmosphere.

The presence of ozone in the stratosphere is essential to our well being on earth. By absorbing a great deal of the lethal ultraviolet radiation from the sun, it makes human life possible on the earth's surface.

1.3 Water vapour in the atmosphere

The atmosphere is never completely dry. Water vapour is always present, but in varying amounts. In hot tropical coastal areas it may reach a concentration of about 3% of the mass of a sample of air. By contrast, in some localities it occurs in such small proportions that it is difficult to measure.

It is remarkable that relatively small amounts of water vapour can produce such great variations in the weather. This is largely due to the changes which occur in its concentration in the troposphere, particularly below about 6 km where a high proportion of the moisture lies. It first enters the atmosphere from the surface of the earth by evaporation from water surfaces and by transpiration from plant life. Water vapour later changes to the liquid or solid state, and finally returns to the earth's surface as dew, frost, drizzle, rain, snow or hail.

On the average, the concentration of water vapour in the atmosphere decreases with altitude. However, from time to time this distribution may be reversed in some parts of the atmosphere.

1.4 Carbon dioxide

Carbon dioxide (CO_2) enters the atmosphere by such processes as human and animal breathing, the decay and burning of materials containing carbon, and volcanic activity. It is largely removed from the atmosphere by plant life.

About 99% of the earth's carbon dioxide is dissolved in the waters of the oceans. However, the solubility varies with the temperature. Hence carbon dioxide enters or leaves the oceans as a result of temperature changes. This affects the concentration of carbon dioxide in the air.

It is estimated that the annual amount of carbon dioxide entering and leaving the air represents about one tenth of the total carbon dioxide content of the atmosphere.

The concentration of carbon dioxide near the ground is variable. In cities near the ground it is high. Away from industrial areas and only a few metres from the soil, departures from the value given in Table 1.1 are relatively small. In the upper atmosphere, however, the distribution of carbon dioxide is less well known.

1.5 The composition of the thermosphere

At altitudes above 80 km the atmospheric gases are less well mixed. There is a tendency for the heavier molecules and atoms to settle out from the others under the action of the force of gravity. This region is known as the thermosphere.

The atoms of many of the gases are also no longer combined together as molecules. This separation is produced by ultraviolet and X-radiation from the sun. For instance, above 80 km oxygen atoms (O) increase with altitude at the expense of diatomic oxygen molecules (O_2). Even at an altitude of 130 km, about two-thirds of the oxygen molecules have been separated into individual oxygen atoms.

As a result of these processes, nitrogen molecules give way to oxygen atoms at high altitudes. Ultimately, these are replaced by light hydrogen atoms at still higher levels.

Short-wave radiation from the sun also produces ionisation. In this process a neutral atom may lose an electron to become a positive ion. By contrast, it may gain an electron and so form a negative ion.

Ionisation may occur below altitudes of 80 km, but the greatest concentrations of ions are found in the thermosphere. In the lower parts of this region, positive and negative ions together with free electrons mix with neutral molecules and atoms. However, at high altitudes protons (ionised hydrogen atoms) and free electrons eventually predominate. At these levels the thin atmosphere of the earth then begins to merge with the very rarefied interplanetary gas.

1.6 **Interplanetary gas**

The space between the planets of the solar system is not a complete vacuum. Although the density of the material is very low, it does contain some hot gases and dust particles.

Material present in these regions in the gaseous state is called interplanetary gas, because it occurs 'between the planets'. It is composed mainly of protons and electrons.

The earth's orbital motion around the sun takes it through the interplanetary gas. In doing so, the outer parts of the earth's atmosphere may be regarded as merging with this very rarefied material.

The atmosphere may be divided into vertical divisions according to its composition, chemical reactions, ionisation, temperature, etc. In the next chapter we shall consider the characteristics of some of these regions.

CHAPTER 2

VERTICAL DIVISIONS OF THE ATMOSPHERE

Meteorologists realise that it is important to consider the atmosphere as a whole, if they are to predict its future behaviour. Increasing use is being made of satellites, rockets and electronic equipment to study the upper atmosphere. At the same time, surface meteorological networks are being extended throughout the world.

It is often convenient to study particular regions of the atmosphere, as we shall do in this chapter. However, it is necessary to keep in mind that events in one region affect those occurring in other parts of the atmosphere.

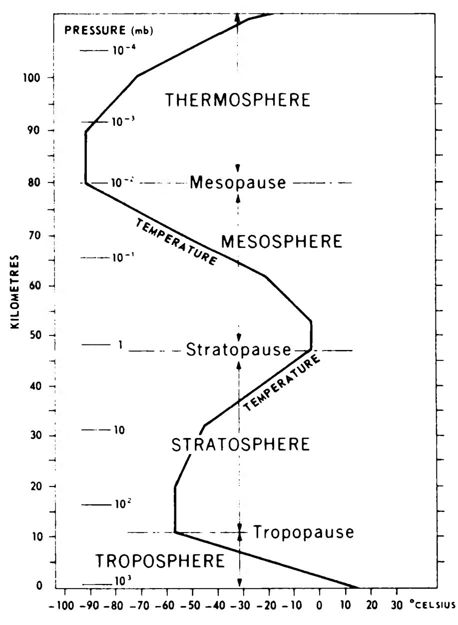
2.1 **Vertical divisions of the atmosphere**

In 1962 the World Meteorological Organization decided to describe the atmosphere by dividing it into four regions—the troposphere, stratosphere, mesosphere and thermosphere. This division was based on the mean variation of temperature with altitude.

Figure 2.1 shows how the temperature and pressure vary with altitude. High temperatures occur near the earth's surface, in the vicinity of the stratopause and in the thermosphere. They are associated with the absorption of radiation from the sun.

Most radiation is absorbed by the earth's surface and so the troposphere is heated

Figure 2.1 Vertical divisions of the atmosphere



from below. By contrast, the heat source in the stratosphere is in its upper levels, where ozone absorbs ultraviolet radiation.

The mesosphere is heated from below. On the other hand, temperatures in the thermosphere begin to rise above the mesopause and continue to do so up to altitudes of at least 300 km, eventually reaching a value of about 1000°C or even higher.

The high temperatures near the stratopause and in the upper thermosphere indicate that the particles are moving at high speeds. However, the density of the atmosphere is very low at these levels.

The number of particles involved is very much greater near the earth's surface. Hence, the heat energy of the atmosphere is largely concentrated in the lower troposphere.

The pressure exerted by the atmospheric gases at any altitude depends on the weight of the particles above unit area at that level. Figure 2.1 shows that the atmospheric pressure near mean sea level is about 1000 mb.

Notice that at the stratopause the pressure is only about 1 mb, indicating that only about one-thousandth of the mass of the atmosphere lies above 50 km. At higher levels the density of the atmosphere becomes even less. Only about one-millionth of the mass of the atmsophere lies above 90 km, and its weight is such that atmospheric pressure is approximately 0.001 mb at that level.

2.2 The troposphere

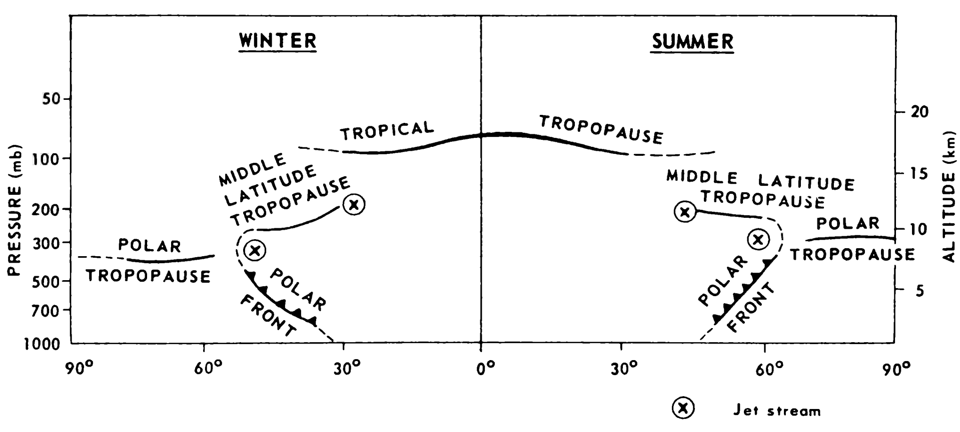
The lowest region is called the troposphere. Usually the temperature decreases with altitude in the troposphere. On the average, the rate of decrease of temperature with altitude is about 6 to 7°C/km in the lower half and 7 to 8°C/km in the upper half.

In some parts of the troposphere there may be a shallow layer in which the temperature increases with height. Such a layer is called a temperature inversion.

The top of the troposphere is called the tropopause. Its altitude varies over the earth and it is not continuous. A tropical tropopause exists in low latitudes at an altitude of about 18 km. At high latitudes a polar tropopause is usually found at approximately 8 km. Between these two regions there is a sloping middle latitude tropopause, with breaks occurring in the vicinity of jet streams. Multiple and overlapping tropopauses occur in middle latitudes and the picture on daily charts is often very complex. Figure 2.2 illustrates the main characteristics of the tropopauses.

In addition, rapid changes in the temperature and altitude of the tropopause may

Figure 2.2 Schematic tropopause patterns



occur in any particular locality. The moving weather systems and the associated clouds that affect our daily lives are almost entirely confined to the troposphere. As these systems change their positions, the characteristics of the tropopause vary in time and place.

The steady decrease of temperature with altitude in the troposphere continues until the tropopause is reached. Because the height of the tropopause is greater over the equator, the lowest temperatures in the atmosphere occur in the vicinity of the equatorial tropopause.

The troposphere contains the greater part of the mass of the atmosphere. It is characteristised by marked vertical motions, appreciable water vapour content, cloud and weather. As a result, it is of great concern to meteorologists.

2.3 The stratosphere

The stratosphere is the region above the troposphere. It extends from the tropopause to altitudes of about 50 to 55 km. In any particular locality, the temperature generally remains practically constant up to about 20 km and this is sometimes called the isothermal layer. The temperature then increases slowly to about 32 km, above which it rises rapidly.

In the upper parts of the stratosphere the temperatures are almost as high as those near the earth's surface. This is due to the fact that ultraviolet radiation from the sun is absorbed by ozone in this region. The low density of the atmosphere at these altitudes means that the solar radiation is transferred to a relatively small number of molecules. As a result, their kinetic energies become large and so the temperature of the air is high.

This heat energy is transferred downwards by subsidence and radiation. The stratosphere therefore has a heat source in its upper levels, in contrast to the troposphere, which is mainly heated from below.

The troposphere and the stratosphere differ greatly in weather conditions. Less convection takes place in the stratosphere, because it is warm at the top and cold at the bottom. It is virtually cloudless, although mother-of-pearl clouds are occasionally observed at high latitudes at altitudes of about 20 to 30 km.

2.4 The mesosphere

At an altitude of approximately 50 km the temperature ceases to rise. This level is the stratopause and marks the lower limit of the mesosphere.

The temperature generally decreases with height in the mesosphere until it reaches a value of −95°C or lower at about 80 km. This level is the top of the mesosphere and is called the mesopause.

The mesopause marks the end of what can be regarded as homogeneous air. Up to this level the gaseous composition of the atmosphere is nearly constant, except for variable amounts of water vapour and ozone. Sometimes the region below the mesopause is known as the homosphere. The homosphere therefore includes the troposphere, the stratosphere and the mesosphere.

The temperatures at the mesopause are lower than at any other level in the upper atmosphere. At high latitudes, noctilucent clouds are occasionally noticed near this level with the sun 5 to 13° below the horizon. It is possible that these clouds consist of dust particles with ice coverings.

2.5 The thermosphere

The thermosphere is the region of rising temperatures above the mesopause. When the sun is quiet this increase may extend to an altitude of about 400 km; during active periods of the sun it may extend still higher to about 500 km.

The composition of the atmosphere changes considerably in the thermosphere. The molecules of many of the gases are separated into their individual atoms by the action of the ultraviolet and X-radiation from the sun.

There is also less tendency for the gases to mix, and the heavier molecules and atoms settle out from the others under the action of gravity. As the altitude increases, the heavier nitrogen molecules give way to oxygen atoms and ultimately light hydrogen atoms predominate at very high levels.

Ionisation is important in the thermosphere, because ions and electrons can remain free for a sufficiently long period of time. It is less persistent in the mesosphere where it normally only exists during the day-time. At the higher pressures in the troposphere, the recombination of positive and electrically charged particles takes place even more easily.

The regions where ionisation is more persistent in the thermosphere and the mesosphere form the ionosphere. The importance of the ionosphere lies in the fact that the electrons in particular are capable of reflecting radio waves.

2.6 The exosphere

The ionosphere, consisting of ions and electrons, extends outwards until it merges with the thin interplanetary gas.

The importance of the neutral gases in the ionosphere should not, however, be overlooked. At an altitude of 160 km, there are still approximately 10^{10} neutral particles, compared with 10^5 electrons in a cubic centimetre of air. Even at about 1200 km the number density of the two components is about equal.

The density of the atmosphere at the mesopause is already very low. In the thermosphere it becomes less and less with increasing altitude. Finally, at an altitude of about 500 to 600 km, the density of the atmosphere becomes so low that collisions between neutral particles become extremely rare. The mean free path is then so large, that escape of the neutral particles from the earth's gravitational attraction becomes possible.

This region is called the exosphere. In the exosphere the neutral atoms and molecules may be regarded as miniature ballistic missiles. Some go up and fall back; some go into orbit around the earth; others escape from the atmosphere and travel into interplanetary space.

The above discussion only applies to the neutral particles of the atmosphere. The motions of the electrically charged particles (i.e. ions and electrons) are controlled by the earth's magnetic field. In fact, even at levels some distance below the exosphere the earth's magnetic field controls the motions of the ions and electrons.

We have mentioned that it is important to study the various constituents of the atmosphere as a whole. However, almost all the clouds and weather that affect our daily lives occur in the troposphere. For this reason, we shall now direct most of our attention to atmospheric processes that occur in the troposphere.

CHAPTER 3

HEAT EXCHANGE PROCESSES IN THE ATMOSPHERE

The enormous amount of energy available to the atmosphere is very apparent during storms. It also becomes evident when broad streams of air sweep across continents and oceans.

Practically all this energy comes from the sun in the form of electromagnetic radiation. The amounts received from the hot interior of the earth and from the stars are quite negligible.

In this chapter we shall study what happens to the solar radiation when it enters the atmosphere. We shall also consider some of the heat exchange processes which occur in the earth-atmosphere system.

3.1 Solar radiation

It has been estimated that approximately 99% of the sun's radiation is contained in the wavelength range of 0.15μ to 4.0μ. Of this radiation 9% occurs in the ultraviolet, 45% in the visible and 46% in the infrared portions of the electromagnetic spectrum. Solar radiation is therefore sometimes known as short-wave radiation.

On the average, only about 43% of the short-wave radiation from the sun is actually absorbed by the earth's surface. Three atmospheric processes—absorption, reflection and scattering—account for the remainder of the radiation.

Most of the ultraviolet radiation is absorbed by ozone in the stratosphere. Water vapour is the only gas to absorb much of the visible radiation. Clouds and dust, however, absorb varying amounts according to the prevailing conditions.

In cloudy conditions a large part of the sun's radiation may be reflected from the tops of clouds and returned to outer space. Some of the radiation reaching the earth's surface is also reflected away.

Solar radiation may also be scattered in all directions by gases and particles in the atmosphere. Some of this scattered radiation is lost to outer space, but some does reach the earth's surface. Solar radiation reaching the earth indirectly is called sky radiation.

The total radiation reaching the earth's surface therefore consists of direct radiation and sky radiation. It is known as global solar radiation.

3.2 Terrestrial radiation

The short-wave radiation from the sun that is absorbed by the earth's surface is converted into heat energy. The average temperature near the earth's surface is about 15°C.

This is, of course, much lower than the temperature of the photosphere of the sun, which is about 6000°C. As a result, the earth radiates long-wave radiation, mainly in the range between $4.0\ \mu$ and $80\ \mu$. The long-wave radiation emitted by the earth and its atmosphere is referred to as terrestrial radiation.

The earth radiates energy most intensely at a wavelength of approximately 10 μ. This radiation is therefore in the form of infrared radiation, rather than visible radiation. In this way it differs from solar radiation, which has a maximum intensity at about 0.5 μ in the visible range.

Substances that absorb only small amounts of solar radiation become significant absorbers and emitters of the long-wave radiation from the earth.

Each atmospheric gas is a selective absorber of terrestrial radiation. It absorbs some wavelengths, but is transparent to others. For instance, ozone absorbs infrared radiation moderately in the wavelength range between 9.6 μ and 15 μ.

Water vapour and carbon dioxide are important absorbers of terrestrial radiation. Between them they absorb most wavelengths of the long-wave radiation from the earth. However, terrestrial radiation can pass through water vapour and carbon dioxide in the wavelength range between about 8 μ and 13 μ. This is known as the 'atmospheric window'.

If clouds are present, they are even more effective absorbers of long-wave radiation. They reflect negligible amounts of terrestrial radiation, in contrast to their substantial reflectivity of solar radiation.

Water vapour, carbon dioxide and clouds in the atmosphere emit long-wave radiation themselves. Some of this energy is returned to the earth's surface, which therefore receives both short-wave radiation from the sun and long-wave radiation from the atmosphere.

If the sky is not overcast, some terrestrial radiation escapes through the 'atmospheric window'. Some of the long-wave radiation absorbed by water vapour, carbon dioxide and clouds is also re-radiated to outer space.

During night-time the incoming solar radiation ceases, but the other processes continue. There is therefore a loss of energy to space, in contrast to the net gain during the day-time.

3.3 **Other heat exchange processes**

The exchange of heat between the earth's surface and the atmosphere does not occur by radiation alone. Conduction and convection are also involved.

In the process of conduction heat passes from the warmer to the colder body without the transfer of matter. It occurs by molecular impacts, the fast moving molecules at the higher temperature collide with and so accelerate the slower molecules at the lower temperature.

Gases are poor conductors of heat. Therefore conduction is only important in heat transfer to extremely thin layers of air which are in direct contact with the earth's surface. These layers are usually only a few centimetres thick and above them heat transfer by conduction is negligible.

Convection is a more important method of transfer of heat energy in the atmosphere. In this process the body carrying the heat itself moves from place to place. In the atmosphere pressure differences arise due to heating of the atmosphere. As a result, the warm air is forced to rise and cold air sinks to lower levels to replace it. Convection currents therefore occur and the air is thoroughly mixed.

Meteorologists make a distinction between sensible heat, which can be sensed or felt, and latent heat, which cannot be sensed directly. Latent heat or 'hidden heat' is the heat added to a substance when it changes its state from solid to liquid or from liquid to gas without change of temperature.

Convection currents in the atmosphere not only transport sensible heat aloft. They also transfer upwards the latent heat stored in water vapour. This latent heat enters the atmosphere when water evaporates from the earth's surface. Later, it is released in the upper air when water vapour condenses to form clouds.

3.4 The energy budget of the atmosphere

The average temperature of the earth has remained approximately constant at about 15°C during the past few centuries. It is therefore in a state of radiative balance, emitting the same amount of energy as it is receiving.

On the average, about 65% of the incoming solar radiation is absorbed by the earth and its atmosphere. This is converted into heat energy and the temperature of the earth's surface and the atmosphere rises.

The radiation from the sun provides the energy for the circulation of the atmosphere and the oceans. This is not lost, but has merely been changed into the form of heat energy or kinetic energy of the particles in motion.

Actually the solar energy may be transformed several times during heat exchange processes between the earth and its atmosphere. These were discussed in the preceding sections.

Eventually, however, the solar energy absorbed by the earth-atmosphere system is re-radiated back to space. By emitting the same amount as it receives, the system maintains its radiative balance.

This balance does not usually apply at individual latitudes. Between latitudes 0 and 35° in both hemispheres more energy is being absorbed than is being radiated out to space, and there is an energy surplus in these regions. Similarly, an energy deficit occurs in regions between latitude 35° and the poles.

Meteorologists have computed the temperatures which would develop if radiative equilibrium were achieved at each latitude without an exchange of heat between different latitudes. A large meridional temperature gradient (from equator to pole) would arise. In actuality, the effective average meridional gradient which is observed is much less. This is because heat is effectively transferred through the latitude circles from low to high latitudes. Both the atmosphere and the oceans are involved in this energy transport.

This meridional transfer of energy is assisted by the action of large-scale eddies (highs and lows), which develop in the regions where strong horizontal temperature gradients occur. Ocean currents also carry some energy away from tropical regions towards the poles.

3.5 The effect of radiation at the earth's surface

When solar radiation reaches the earth's surface, it is affected in various ways, ranging from almost total reflection to practically complete absorption. These effects depend largely on the nature of the surface receiving the radiation.

The albedo of a surface is defined as the ratio of the amount of global radiation from the sun and sky reflected by the surface to that which falls on it, i.e.

$$\text{albedo (of surface)} = \frac{\text{global radiation reflected by surface}}{\text{global radiation falling on surface}}$$

A great deal of the radiation falling on snow is reflected. The albedo of snow surfaces ranges from over 0.80 for cold, fresh snow to about 0.50 for old, dirty snow surfaces.

The more common land surfaces, such as forests, grasslands, ploughed fields and rocky deserts, have an albedo ranging from about 0.10 to 0.20. On the other hand, the values may range from 0.30 for dry sand down to about 0.05 for dark forests.

Water absorbs a large proportion of the incoming radiation if the angle of elevation of the sun is high. By contrast, it reflects most of the radiation if the sun is low in the sky.

3.6 **Temperature differences between land and sea surfaces**

The rise in temperature of the earth's surface when it absorbs radiation varies. It depends partly on the distance to which heat penetrates and partly on the specific heat of the material.

The specific heat of a substance is the quantity of heat required to raise the temperature of unit mass of the substance by 1° C. With the exception of hydrogen, water has the highest specific heat of any substance. It requires a relatively large amount of heat energy to raise the temperature of unit mass of water by 1°C.

Sand, depending on its colour, absorbs different amounts of radiation. Its specific heat is low and so its temperature rises rapidly when it is heated. In addition, it is a poor conductor and only a thin layer of sand absorbs the radiation. As a result, the temperature of a sand surface rises rapidly during the day-time.

During the night-time, incoming solar radiation ceases and the sand loses heat by radiation. It therefore becomes increasingly colder during the night. Sand surfaces are therefore subject to large extremes of temperature between day-time and night-time. Similar effects occur when insolation falls on rock and other land surfaces.

Water absorbs a large proportion of the incoming radiation when the sun is high in the sky. However, it has a high specific heat and so its temperature only rises slowly. Some of the radiation penetrates the water to a depth of several metres, while mixing of the surface layers tends to spread any temperature changes through a considerable depth. In addition, some of the heat energy gained by the water is converted into latent heat during the evaporation process.

The temperature of the sea therefore does not rise as rapidly as that of a land surface during the day-time. At night, the incoming radiation ceases and heat is lost by radiation. However, there is usually a large store of heat energy below the water surface, and so very little change occurs in the surface temperature. The variation in the temperature of sea surfaces between day-time and night-time is therefore very small.

The temperature of the earth's surface indirectly affects the temperature of the gases of the atmosphere. In our daily lives we are concerned with the temperature of the air itself and we shall consider this aspect in the next chapter.

CHAPTER 4

AIR TEMPERATURE

The idea of temperature was originally conceived on the basis of feeling. An object was judged to be hot or cold according to how it felt to the touch.

The temperature of a body is the condition which determines its ability to transfer heat to other bodies or to receive heat from them. In a system of two bodies, the one which loses heat to the other is said to be at a higher temperature.

In this chapter we shall be primarily concerned with the temperature of the air. We shall first deal with its measurement and then later investigate its variation in time and space.

4.1 Basic principles of temperature measurement

With the advance of scientific method accurate measurements of temperature became necessary. It was realised that as the temperature of an object increased, certain physical changes occurred. These included the expansion of solids, liquids and gases. Changes of state also occurred—solids melted and liquids boiled.

A thermometer is a temperature-measuring instrument. Many physical properties of matter are utilised in thermometers. These include the expansion of solids, liquids and gases and the change of electrical resistance with temperature.

Instruments used for measuring high temperatures are called pyrometers. For instance, radiation pyrometers depend on the heat radiation emitted by a hot body. These have the advantage that they are not in contact with the hot body whose temperature is being measured.

It should be realised, however, that temperature has no dimensions and cannot be measured in the same way as, say, length. Thus, if an object is 15 m long, we can place fifteen units of length along it from one end to the other. In the case of temperature there are no units.

We simply specify two fixed points, the temperatures of which can be reproduced as specified physical conditions of certain substances. Numbers are then assigned to the temperatures of these fixed points and so a specified number of divisions occur between the fixed points of the temperature scale.

We therefore speak of temperature division and not temperature units. Thus 20°C is not twice the temperature of 10°C; 10°C and 20°C correspond with the tenth and twentieth divisions of the Celsius scale between the fixed points corresponding to 0°C and 100°C.

4.2 Celsius and Fahrenheit temperature scales

Practical temperature scales are based on fixed points. These are constant and easily reproducible temperatures. Two internationally-agreed fixed points are the ice point and the steam point.

The ice point is the temperature at which pure ice melts under an external pressure of one standard or normal atmosphere. This is the pressure that will support a column of mercury 76.0 cm high and is equal to 1013.250 mb. (The ice point is related to the fundamental fixed point, known as the triple point of pure water, which will be discussed in Section 4.4.)

The steam point is the temperature at which pure water boils under an external pressure of one standard atmosphere.

Two temperature scales are in common public use. One is the Celsius (or Centigrade) scale on which the ice point is 0°C and the steam point 100°C.

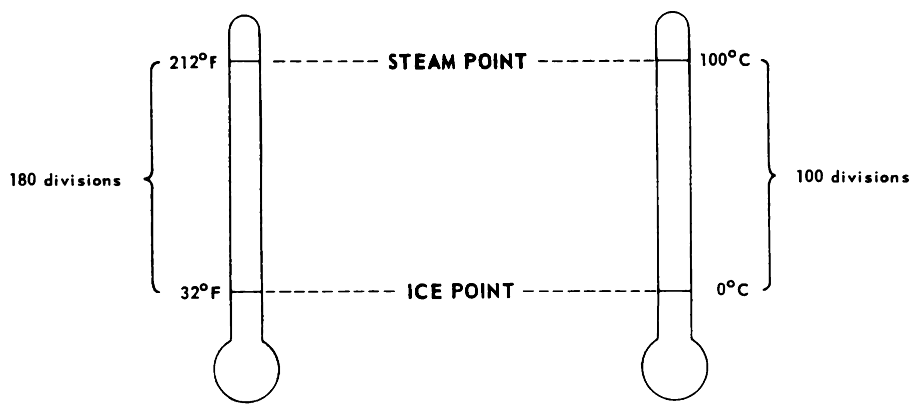
The other scale is the Fahrenheit scale. On this scale the ice point is labelled 32°F and the steam point 212°F.

Notice that there are 180 Fahrenheit divisions between the ice and steam points compared with 100 Celsius divisions in the same range. Thus there are 180/100 or 9/5 Fahrenheit divisions for each Celsius division.

The number allotted to the ice point on the Fahrenheit scale is also greater by 32 than that on the Celsius scale.

The relationship between the two scales is shown in Figure 4.1.

Figure 4.1 Temperature scales



4.3 **The Kelvin scale of temperature**

In scientific work, the Kelvin scale of temperature is frequently used. It is related to the Celsius scale of temperature by the following formula:

$$K = 273.15 + °C \qquad \ldots 4.3$$

where K = temperature in kelvins.

For instance, 20°C is equivalent to 293.15 K.

The Kelvin scale of temperature is often referred to as the absolute scale of temperature.

The fundamental fixed point of the Kelvin scale of temperature is the triple point of pure water. This is the temperature at which the solid, liquid and vapour states of pure water occur in equilibrium together. It is assigned the temperature 273.16 K and is therefore 0.01 K higher than the ice point.

4.4 Physical processes used in thermometry

There are many types of thermometers in use. Some of the principles employed are based on the following effects of increasing temperature:

(a) Expansion of liquid in a glass container
(b) Expansion of liquid in a sealed container causing an increase in pressure
(c) Voltage generated at the junction of two dissimilar metals (thermocouple)
(d) Increased twist of a metallic strip which consists of two different metals fastened together along their length (bimetallic thermometer)
(e) Change in resistance of a platinum wire
(f) Change in resistance of a special mixture of chemicals (thermistor).

Some of these effects are also utilised in thermographs. A thermograph is a self-recording thermometer which makes a continuous record of temperature measurements.

4.5 Principles of design of thermometers

Details of the design of some of the more important types of thermometers are as follows.

Liquid-in-glass thermometers

The liquids used are usually either mercury or ethyl alcohol. Mercury can only be used down to about -36°C, which is just above its freezing point. For lower temperatures, absolute ethyl alcohol is generally suitable.

The thermometer consists of a glass tube of small bore through which the liquid rises from the bulb, which may be spherical or cylindrical. The top of the liquid column indicates the temperature.

Special types of liquid-in-glass thermometers are made to show either the highest (maximum) temperature or the lowest (minimum) temperature.

Liquid-in-metal thermometers

The indicating unit of these thermometers is really a pressure gauge calibrated to read temperature. It is widely used in motor car engines as a temperature gauge.

This principle is also used in some thermographs. A pen is fitted to the tip of the pointer and the nib moves across a chart placed on a rotating cylindrical drum.

Thermocouples

A thermocouple consists of two wires of different metals joined at each end. If a difference in temperature develops between the two junctions a thermo-electric e.m.f. is generated. A specially calibrated voltmeter indicates the difference in temperature between the two junctions.

Thermocouples are widely used as pyrometers, i.e. for measuring high temperatures. In specialised applications they are also used for low temperature measurement.

Bimetallic thermometers

Two metals are selected which have markedly different rates of expansion with temperature change. One or both of these may be alloys.

These consist of strips which are strongly bonded together on their flat surfaces. When the temperature changes, one strip becomes longer than the other. This causes the combined strip to bend with the longer metal on the outside of the curve.

The strip may be wound in the form of a coil. The inner strip is usually made of the metal which expands more rapidly. As a result the coil unwinds when the temperature rises. This movement is magnified by a simple lever system attached to the coil, and a pointer indicates the temperature.

This principle is commonly used in thermographs to obtain a continuous record of the temperature.

Platinum resistance thermometers

The principle of operation of this instrument is based on the change in resistance of a platinum wire when the temperature is altered. A battery provides a source of electric current to operate a resistance meter which is calibrated to read temperature. The instrument can also be designed to produce a continuous record of temperature and so act as a thermograph.

The platinum resistance thermometer is an extremely accurate instrument. It is capable of measuring a wide range of temperatures.

Thermistors

There are certain chemical substances which change their electrical resistance markedly with temperature, the resistance decreasing with rising temperature.

Thermistors are small and robust. They are often used as radiosonde thermometers, which are carried aloft by balloons to obtain upper air temperatures. Variations in the resistance of an electrical circuit occur as the temperature changes with altitude. These alter the radio signals transmitted to a receiver on the earth's surface. The changes are recorded on a chart and the air temperature is determined at various levels up to about 30 km.

4.6 **Measurement of air temperature**

A thermometer (or thermograph) indicates the temperature of its heated element. This may be some temperature other than that of the air whose temperature we wish to determine. For instance, radiant heat passes through the air without affecting its temperature, but is absorbed by the material of the thermometer.

It is also essential to ensure that the air reaching the thermometer is at the same temperature as that of the air whose temperature is being measured. Air can be heated by contact just before it reaches the thermometer. A false reading is then obtained.

4.7 **Surface air temperature**

In meteorological practice, the surface air temperature refers to free air at a height between 1.25 and 2 m above ground level. In general, it is found that this temperature is representative of the conditions experienced by human beings living on the earth's surface.

The surface air temperature measured in this way can, however, differ considerably from the temperature of the ground below. On hot sunny days the ground temperature may be somewhat higher than that of the surface air temperature. By contrast, it can be much lower than the surface air temperature on cold, frosty nights.

4.8 **Thermometer exposure**

To give a representative reading of air temperature, thermometers should be protected from radiation from the sun, sky, earth and any surrounding objects. At the same time, they should be adequately ventilated so that they indicate the temperature of the free air.

Two methods of protection are in general use. These are:

(a) The louvered thermometer screen
(b) Polished metal shields with forced ventilation.

In either case, the equipment should be installed in such a position as to ensure that the measurements are representative of the free air circulating in the locality. The temperature should not be influenced by artificial conditions, such as large buildings and expanses of concrete or tarmac.

As far as possible, the soil cover beneath the instruments should be short grass or, at places where grass does not grow, the natural earth surface of the district.

Details of the meteorological requirements for thermometer screens and shielded ventilated thermometers are given in the WMO *Guide to Meteorological Instrument and Observing Practices* (WMO—No. 8 TP.3).

4.9 **Diurnal variation of surface air temperature**

During the twenty-four hours of each day, temperature changes are much more gradual over the sea than over the land. The diurnal variation in sea-surface temperature is usually less than 1°C, and the air temperature near the water surface is equally steady in quiet conditions.

On the other hand, in desert regions in the interior of continents surface air temperatures may vary from day to night by as much as 20°C. Near the coast, however, the diurnal variation of temperature depends largely on the direction of the wind, being large if the wind is off the land and small if it is from the sea. Local land and sea breezes also tend to reduce the range of temperature variation.

In general, the diurnal variation of surface air temperature tends to be greatest if calm conditions prevail. If it is windy, mixing of the air occurs through a deeper layer. The gain of heat by day and the loss by night is then shared by more molecules of the atmospheric gases. As a result, the diurnal range of temperature may be reduced during windy conditions.

Cloudiness reduces the diurnal range of temperature at any place. During the day-time clouds only absorb or transmit a small amount of the radiation from the sun. The major part is reflected back to space and does not reach the earth's surface.

By contrast, at night clouds absorb the long-wave radiation being radiated upwards from the earth's surface. They then re-radiate most of this heat energy back to the earth's surface again. In this way, they act like a blanket keeping the earth's surface warm. Diurnal variation of surface air temperatures is therefore relatively small during cloudy conditions.

The type of surface and the ability of the underlying material to conduct heat to or from the surface affect the diurnal range of surface air temperature. However, the nature of the neighbouring terrain is also important, because the temperature at a particular place may be affected by the flow of warm or cold air from adjacent areas.

For instance, surface cooling takes place at night due to radiation. Air near the ground then cools and becomes denser. If the surface slopes, the cooled air flows to lower levels as a katabatic wind. The opposite effect occurs during the day-time when an anabatic wind flows up the slope. Warm air on the heated slope is replaced by cooler, denser air from lower levels. However, an anabatic wind is generally not as well developed as a katabatic wind, due to the fact that its motion is reduced by the force of gravity.

The effects of surroundings are evident in large cities. On still, clear nights the temperatures in the centre of a city may be more than 5°C higher than those in nearby playing fields. Day-time temperatures are also affected by (heat generated by) activities in the city buildings.

4.10 **Temperature variation in the vertical**

We noticed in Section 2.2 that, in general, the temperature decreases with altitude in the troposphere. The rate of decrease of temperature with altitude is called the temperature lapse rate.

On the average, the temperature lapse rate in the troposphere is about 6°C per km. This means that, if the temperature at mean sea level is 15°C, it will decrease to a value of about −15°C at 5 km (i.e. a fall of 30°C).

In the lower levels of the stratosphere the temperature sometimes does not change with altitude. The temperature lapse rate is then zero. The atmosphere is said to be isothermal in such regions. Isothermal means 'equal temperature'.

In some parts of the atmosphere the temperature increases with altitude. A negative temperature lapse rate then occurs. Remember that a negative lapse rate indicates a rise in temperature with altitude. For instance, suppose the temperature increases by 2°C during an altitude increase of 1 km. We then say that the temperature lapse rate is −2°C per km.

Normally, the temperature decreases with altitude in the troposphere. The temperature lapse rate is on the average positive with a value of about 6°C per km.

However, sometimes the temperature increases with altitude for some distance in the vertical. A temperature inversion is then said to occur, i.e. the normal change of temperature in the troposphere has been inverted or reversed.

If you refer to Figure 2.1 again you will notice that, on the average, a temperature inversion also occurs in the upper part of the stratosphere. By contrast the temperature in the mesosphere decreases with altitude on the average, i.e. the temperature lapse rate is positive.

In the thermosphere the temperature increases with altitude and so the lapse rate of temperature becomes negative in this region.

The temperature of the atmosphere affects its density. This determines the weight of the air which lies above a surface of unit area near the earth's surface, i.e. the pressure. In the next chapter we shall study some of the characteristics of the atmospheric pressure.

CHAPTER 5

ATMOSPHERIC PRESSURE

As we go about our daily lives we are subjected to the pressure produced by the weight of the gases of the atmosphere above. Billions and billions of molecules and atoms move at high speed around and above us colliding with one another, with the earth's surface, with human beings and with any object that crosses their paths.

The study of atmospheric pressure forms one of the basic sections of meteorology. Differences of pressure within the atmosphere lead to the flow of air from one place to another. Winds and ultimately all the various elements of the weather arise in this way.

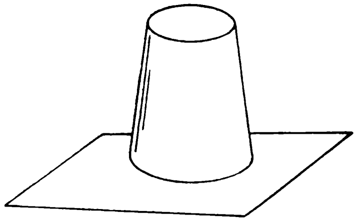
5.1 The nature of atmospheric pressure

Scientists make a distinction between force and pressure. Pressure is the force on a unit of area.

The high-speed molecules and atoms of nitrogen, oxygen and other atmospheric gases bombard any surface area which comes in contact with them. The force which they exert on unit area of that surface is called the atmospheric pressure.

Because the gases are moving in all directions, they exert a pressure on any surface, no matter which direction it faces. The fact that air exerts a pressure in all directions may be shown by filling a tumbler to overflowing and placing a sheet of light cardboard over its mouth. When the tumbler is carefully inverted (see Figure 5.1) it will be found that the cardboard is firmly held on the mouth of the tumbler.

Figure 5.1 Air exerts a pressure in all directions



The air pressure is greater than the downward pressure due to the weight of the water. By slowly turning the tumbler to face in various directions, it can be shown that air pressure acts sidewards as well as upwards.

At any particular locality the pressure is greatest near the earth's surface, as it is produced by the weight of the whole of the air above any unit surface area. Higher in the atmosphere there are fewer molecules and atoms above, and so the atmospheric pressure decreases with altitude.

5.2 Units of atmospheric pressure

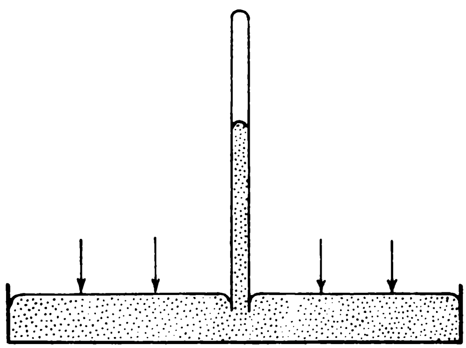
Near the earth's surface, the air exerts a pressure of about 10^5 newtons on each square metre of surface. This pressure is equal to one bar.

Small variations in pressure occur during the day, and it is therefore desirable to use a smaller unit, so that these changes may be reported. The unit used for meteorological measurements is one-thousandth of a bar. It is called the millibar, i.e. 1 bar = 1000 mb.

A millibar is therefore equivalent to a force of 100 newtons on each square metre of surface in contact with the air.

The pressure of the air is able to support the weight of a column of mercury enclosed in a tube which is free of air above the mercury surface. Figure 5.2 shows this effect.

Figure 5.2 Air pressure supporting the weight of a mercury column



The length of the mercury column will of course vary with the temperature, while its weight will depend on the force of gravity at a given locality. Meteorologists therefore determine the length of the mercury column under standard conditions. Under standard conditions the temperature is that of melting ice (0°C) and the acceleration due to gravity is 9.80665 m/s^2.

If the air is able to support the weight of a column of mercury equal in length to 0.760 m (or 760 mm) under standard conditions, the pressure is called a standard atmosphere. It is equivalent to 1013.250 mb.

Notice that, under standard conditions, if

760 mm of mercury = 1013.250 mb
then 1 mm of mercury = 1.333224 mb.

This is known as a standard millimetre of mercury.

5.3 The measurement of atmospheric pressure

An instrument used to measure atmospheric pressure is known as a barometer. The word barometer is derived from two Greek words *baros* (weight) and *metron* (measure).

There are two types of barometer in general use—mercury and aneroid. Aneroid is derived from the Greek *a nexos* (not wet) and the suffix *oid* (like). An aneroid barometer therefore has no fluid, in contrast to a mercury barometer, which uses a liquid.

5.4 Mercury barometers

In 1643 the Italian scientist Toricelli took a glass tube about 80 cm long and closed at one end. He filled it with mercury and inverted the open end under the surface of mercury in a bowl.

Toricelli found that the mercury level in the tube fell to and remained about 0.76 m (or 760 mm) above the outside mercury surface. He explained this behaviour by stating that the atmosphere must be exerting a pressure on the free mercury surface. This pressure must be equal to the pressure produced by the weight of the mercury column.

This was the first mercury barometer. The vertical height of the column of mercury does not depend on the slope of the tube. If it is measured, the atmospheric pressure can be stated as being equivalent to a certain length of mercury. This length must, however, be corrected, so that it refers to standard conditions of temperature and acceleration due to gravity.

Two types of mercury barometers are used at meteorological stations. They are the Fortin and the fixed cistern (or Kew pattern) barometers.

The length to be measured is the distance between the top of the mercury column and the upper surface of mercury in the cistern. However, a difficulty arises due to the fact that any change in the length of the mercury column leads to a change in the level of the mercury in the cistern.

In the Fortin barometer the level of the mercury in the cistern can be adjusted to bring it into contact with a fixed ivory pointer. The top of this pointer is at the zero of the barometer scale. The first operation in reading a Fortin barometer is therefore to adjust the level of the mercury in the cistern until it corresponds to this point.

The fixed cistern barometer is often called the Kew pattern barometer. In using the instrument it is not necessary to adjust the level of the mercury in the cistern. The scale engraved on the barometer is contracted to allow for changes in the mercury level.

5.5 Corrections for standard conditions

The length of a mercury column of a barometer depends on other factors (especially temperature and gravity) in addition to the atmospheric pressure. It is therefore necessary to specify the standard conditions, under which the barometer should theoretically yield true pressure readings. Scales on mercury barometers for meteorological purposes should be so graduated that they yield true pressure readings directly in standard units under standard conditions. This applies when the entire instrument is maintained at the standard temperature of 0°C and at the standard value of gravity of 9.80665 m/s^2.

In order that barometer readings made at different times and at different places should be comparable, the following corrections should be made:

Correction for index error
Correction for temperature
Correction for gravity.

Correction for index error

A barometer is graduated on the assumption that a given reading on the scale represents the actual difference between the levels of the mercury surfaces. In practice, it is frequently impossible to position or subdivide the scale perfectly.

In addition, glass is not wetted by mercury and the meniscus (i.e. curved surface of the liquid in the tube) is convex. The force of cohesion between the mercury molecules is greater than that of adhesion between the glass and the mercury.

As a result, the mercury level in the tube is depressed. This movement is in the opposite direction to the rise of water in thin tubes. In each case, the effect is known as capillarity.

Small errors also arise due to residual gas in the vacuum above the mercury column. In addition, refraction or bending of the light rays occurs as they pass through glass.

These various errors should not exceed a few tenths of a millibar in good mercury barometers. They are combined together to form the index error indicated on the calibration certificate for the barometer. This is prepared after comparison tests have been made with a standard barometer.

Correction for temperature

Barometer readings have to be corrected to the value that would have been obtained if the mercury and the scale had been at the standard temperature of 0°C.

Mercury barometers used for meteorological purposes are calibrated by comparison with accurate standard barometers. When the barometer is initially calibrated the various component parts, such as the mercury, scales, cistern, glass tube, etc., are at a temperature of 0°C. Any departure from this temperature will cause changes in the dimensions of these components.

For this reason, a thermometer is fixed to mercurial barometers in such a position that it gives the mean temperature of the component parts. It is called the attached thermometer. By reading this thermometer it is possible to make a correction to the barometer measurement so that it will refer to the standard temperature of 0°C.

Correction for gravity

The reading of a mercurial barometer at a given pressure and temperature depends on the value of the acceleration due to gravity. This in turn varies with latitude and with altitude.

Barometers are calibrated to yield true pressure readings at the standard gravity acceleration of 9.80665 m/s^2. Their readings at any other value must be corrected.

For any particular barometer used in a fixed position these corrections may be conveniently combined in a single table. The station (level) pressure is then obtained by applying the single correction, which corresponds to the temperature of the attached thermometer.

5.6 Aneroid barometers

An aneroid barometer consists of a flexible sealed metal chamber. It is completely or partly evacuated and the distance between the centres of its opposite walls changes with atmospheric pressure. For instance, if the atmospheric pressure increases, it forces the opposing walls together.

A strong spring system prevents the chamber from collapsing due to the external atmospheric pressure. At any given pressure there will be an equilibrium position, in which the force due to the spring balances that of the external pressure.

One end of the cell is fixed, while the other is coupled to a pointer which moves over a dial marked with pressure values. The coupling magnifies the movement of the pointer.

An aneroid barometer must be calibrated against a mercury barometer. However, it has a great advantage over the mercury barometer in that it is compact and portable. An aneroid barometer is therefore very convenient to use at sea or in the field.

Sources of error in aneroid barometers include incomplete compensation for temperature. A weakening of the spring may occur with increasing temperature. This will result in too high a pressure being indicated by the instrument.

Elasticity errors also occur. When an aneroid barometer is subjected to a large and rapid change of temperature the instrument will not immediately indicate the true pressure. This lag is known as hysteresis, and a considerable time may elapse before this difference becomes negligible.

Slow changes in the metal of the aneroid chamber also occur. These are known as secular changes and can only be allowed for by comparisons at intervals with a standard barometer.

5.7 **Barographs**

A barograph is a self-recording barometer, which makes a continuous record of pressure measurements over a period of time. It usually employs an aneroid mechanism.

A number of aneroid cells are fixed together, so that there is more force to move the pointer. By a suitable system of levers, the expansion or contraction of the cells is magnified. This movement is transmitted to a pointer, which moves in an arc over a paper chart wrapped around the outside of a cylindrical drum. The drum is rotated once weekly by a clockwork mechanism and so a continuous record of the atmospheric pressure at the meteorological station is obtained.

5.8 **Variation of pressure with altitude**

The pressure of the atmosphere at the earth's surface represents the weight of a column of air of unit area of cross-section, extending from the earth's surface to the upper limits of the atmosphere. At higher altitudes the pressure is lower, due to the fact that there is less air above any level than there is above the earth's surface.

For instance, the pressure near the earth's surface is about 1000 mb, whereas at an altitude of about 5.5 km it is only half this value. Table 5.1 indicates the mean pressure at higher altitudes in middle latitudes.

Table 5.1 Variation of pressure with altitude

Altitude (km)	Pressure (mb)
10	265
20	55.3
30	12.0
40	2.87
50	0.798
60	0.225
70	0.0552
80	0.0104
90	0.0016
100	0.0003

The rate at which the pressure decreases with altitude is not constant. For example, near mean sea level, the pressure decreases by one mb for a rise of about 8.5 m. At about 5.5 km, the same pressure decrease is equivalent to an altitude interval of approximately 15 m. With increasing altitude the interval becomes still greater. Only approximate values can be given, however, because the temperature affects the decrease of pressure with altitude.

5.9 **Reduction of pressure to standard levels**

The value of the pressure at the observing station is called the station (level) pressure. To enable barometer readings at different elevations to be compared, it is first necessary to reduce them to the same level.

In most countries the observed atmospheric pressure is reduced to mean sea level. The pressure obtained in this way is known as the mean sea level pressure.

Mean sea level pressures must be hypothetical in most cases. The stations are on land, and so one has to imagine that a column of air occupies the space occupied by the earth and rock below the station. This column of air is then assumed to extend down to mean sea level.

To compute mean sea level pressure, it is first necessary to determine the station (level) pressure. We then add to this value the weight of air which would occupy a column of unit area of cross-section between the station and mean sea level.

The height of this column is fixed, but its weight depends on its density and therefore the temperature. If the temperature rises, the air becomes less dense. The weight to be added is then reduced.

There is no really satisfactory way of reducing elevated inland stations to mean sea level. It is impossible to decide what the mean temperature of the hypothetical column of air should be.

Some countries use the air temperature at the station itself. This gives satisfactory results for stations of elevations below 500 m.

The World Meteorological Organization has recommended a method for use at low-level stations in the *Guide to Meteorological Instrument and Observing Practices* (WMO—No. 8 TP.3).

In spite of the difficulties involved, reduction of station level pressure to mean sea level can be satisfactorily carried out in many regions. Mean sea level synoptic pressure charts are widely used by meteorologists throughout the world. The word synoptic is derived from the Greek words *syn* (together) and *opsis* (seeing). Pressure readings are taken simultaneously at a large number of stations, corrected to mean sea level and then plotted on synoptic charts.

In parts of Africa and Antarctica the elevation of meteorological stations is well above 1 km. These stations, after obtaining station (level) pressure, compute the approximate altitude at which 850 mb or 700 mb pressures occur above them. Synoptic weather charts are then drawn at these pressure levels.

5.10 **Altimetry**

We have noticed that there is a close connection between atmospheric pressure and altitude. Use is made of this effect in aviation to determine the altitude at which an aeroplane is flying.

The pressure altimeter is an aneroid barometer with its scale graduated to read altitude instead of pressure.

We have seen that the pressure at the earth's surface is equal to the weight of a column of air of unit cross-section, extending to the outer limits of the atmosphere. The pressure therefore depends on the density of the gases in the atmosphere. In turn, the density is affected by the temperature of the gases. The higher the temperature, the lower the density.

Variations in the temperature with altitude therefore introduce difficulties in altimetry. Altimetry deals with the measurement of altitude by means of aneroid barometers.

For this reason, it is desirable to first assume that a hypothetical atmosphere exists in which certain variations in temperature occur with height. This is known as the ICAO standard atmosphere and will be discussed in the next section.

In practice, significant differences occur between the actual atmosphere and this hypothetical standard atmosphere. However, allowances can be made when the departures from the standard atmosphere are known.

5.11 **The ICAO standard atmosphere**

The International Civil Aviation Organisation (ICAO) has established standard conditions on which the scale of an altimeter can be based. This is known as the ICAO standard atmosphere. Some of the details are shown in Table 5.2.

It is assumed that the air is dry. The specifications show that the mean sea level pressure and temperature are 1013.25 mb and 15°C respectively. A lapse rate of 6.5°C per kilometre is assumed up to a tropopause at about 11 km.

The atmosphere is then isothermal in the lower stratosphere, until an altitude of approximately 20 km is reached. Thereafter, the lapse rate is assumed to be negative, and the temperature rises at the rate of 1.0°C for every increase in altitude equal to 1 km. For aviation purposes, it is not necessary to specify the standard atmosphere above 32 km.

Table 5.2 Some specifications of the standard atmosphere

Approximate altitude (km)	Pressure (mb)	Temperature (°C)	Temperature lapse rate (°C per km)
0	1013.25	15	6.5
11	226.32	−56.5	0.0
20	54.75	−56.5	
32	8.68	−44.5	−1.0

5.12 **Altimeter adjustments**

Corrections must be made to pressure altimeter readings whenever actual conditions differ from those specified for the ICAO standard atmosphere.

The altimeter is constructed so that the altitude and pressure scales can be rotated relative to each other. If the actual mean sea level pressure differs from 1013.25 mb. an adjustment can be made. The altimeter will then read zero altitude at the existing mean sea level pressure.

No adjustments to the altimeter are made for departures of actual temperature values from standard conditions. However, estimates can be made from the temperature at flight level and surface temperature. For instance, if the temperature is lower than for the standard atmosphere, the density of the air will be above average, and so the altimeter will read too high. If, however, the observed air temperature is set against the indicated altitude on a navigational computer, the corrected value may be determined.

Further details on the use of pressure altimeters are provided in the *Manual of Meteorology Part 2 Aviation Meteorology.*

5.13 The semi-diurnal variation of pressure

The atmosphere at a given locality varies continually. These variations may be regular or irregular.

Irregular variations are sometimes due to the passage of pressure systems. The development or decay of a pressure system may also produce changes of this type.

The regular variations have various periods. The most important regular oscillation has a period of about twelve hours (or ½ day). It is therefore called the semi-diurnal variation of pressure.

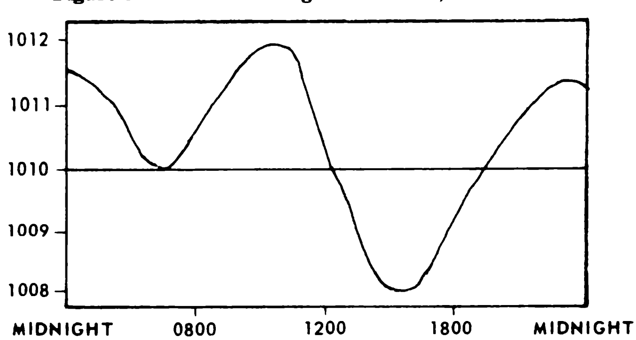
Radiation from the sun leads to the alternate heating and cooling of the earth's atmosphere. This produces a rhythmic expansion and contraction of the atmosphere as the earth rotates and as a result, pressure oscillations occur.

The atmosphere itself is assumed to possess a natural period of oscillation of about 12 hours. This is excited by the temperature variations, and the amplitude of the oscillation increases due to resonance. As a result, a double wave travels around the earth following the sun. Pressure maxima occur around 1000 and 2200 local mean time, with minima at about 0400 and 1600.

The semi-diurnal variation of pressure is rather complex. The changes are not perfectly symmetrical and they vary considerably with locality. The variations have little influence on other meteorological factors, but must be kept in mind when interpreting pressure changes.

In the tropics, the semi-diurnal variation of pressure is more evident than in higher latitudes. Figure 5.3 shows the pressure variations for a typical day at Darwin, Australia.

Figure 5.3 Pressure changes at Darwin, Australia



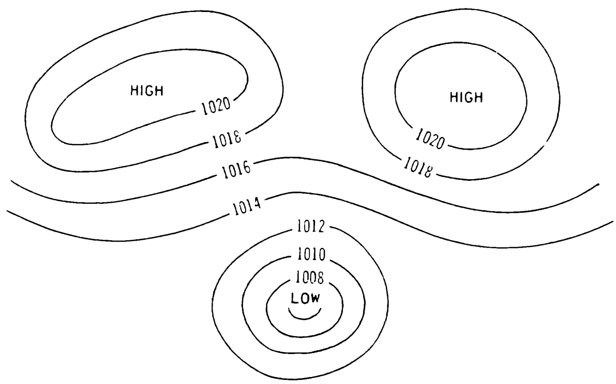
In middle and high latitudes, the semi-diurnal variation of pressure is sometimes difficult to detect, due to the frequent passage of pressure systems. However, the regular variations can be computed by averaging the pressures at each hour for a long period of time, so that the pressure changes due to storms are eliminated.

5.14 **Pressure gradient**

When simultaneous pressure readings from a large number of stations are reduced to mean sea level, they are plotted on a mean sea level synoptic chart. Meteorologists then draw isobars on these charts. An isobar is a line joining places having the same pressure.

Some isobars surround areas of high pressure and others enclose regions of low pressure. This is shown on Figure 5.4. In the example the difference in pressure between adjacent isobars is 2 mb.

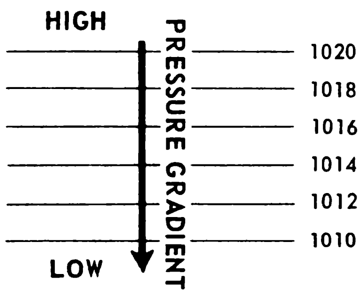
Figure 5.4 Mean sea level synoptic pressure chart



In some regions the isobars are very close together, indicating that as one crosses the isobars from high to low pressure, rapid changes in pressure occur. This may be compared with motion down a steep slope or gradient from high to low levels.

Pressure gradient is the change of pressure with horizontal distance measured in the direction from high to low pressure. At any point the pressure gradient is directed at right angles to the isobar. This is indicated in Figure 5.5.

Figure 5.5 Pressure gradient



The pressure gradient is described as 'steep' or 'strong' if the isobars are closely spaced. By contrast, it is considered to be 'flat' or 'weak' when the isobars are far apart.

In a later chapter, we shall see that the force exerted on the gases of the atmosphere as a result of the pressure gradient is an important factor in determining the strength of the wind. The weather at any locality is directly affected by the difference between its pressure and that of other adjacent stations.

However, before we consider these relationships, it is necessary to study the moisture content of the air. In the next chapter we shall therefore consider the humidity of the atmosphere in some detail.

CHAPTER 6

MOIST AIR

Although water is present in some degree everywhere in the atmosphere, it usually occurs in the invisible state in the form of vapour. From time to time, however, it condenses to form clouds, which provide some evidence of the weather to come.

Water enters the atmosphere by the processes of evaporation and transpiration. Later, it returns to the earth as precipitation and so completes the water cycle.

If we are to understand these processes and so predict the future state of the atmosphere, we must consider the variations that occur in the humidity or moisture content in the atmosphere. We must also investigate the methods used to measure the humidity of the air.

6.1 **Moist air**

In addition to the dry-air constituents listed in Table 1.1, the atmosphere contains a variable amount of water vapour.

Air consisting of dry air and water vapour is called moist air. Before we study the characteristics of moist air, we shall consider some of the processes by which water changes from one state to another.

6.2 **The three states of water**

The various materials which comprise the earth occur in three different states—solid, liquid and gas. The three states of water are as follows:

Solid state—ice
Liquid state—water
Gaseous state—water vapour.

Water may change from one state to another either directly or indirectly. The processes by which water changes from one state to another are as follows:

Initial state	Final state	Process
Ice	Water	Melting
Ice	Water vapour	Sublimation
Water	Water vapour	Evaporation
Water vapour	Water	Condensation
Water vapour	Ice	Deposition
Water	Ice	Freezing

6.3 **The vapour pressure of moist air**

The atmosphere is a mixture of gases, each of which exerts its own pressure, called its partial pressure. The partial pressure exerted by each gas is proportional to the number of molecules of that gas present in the given volume of the gas mixture.

The atmospheric pressure at any point is equal to the sum of the partial pressures exerted by each of the gases in the atmosphere (including water vapour).

As water evaporates into dry air, the vapour begins to exert its own partial pressure—the water vapour pressure (e). The atmospheric pressure (p) is increased, because it is now equal to the sum of the pressures exerted by both the water vapour and the dry air.

6.4 Variations in water vapour content of the atmosphere

The quantity of water vapour present in the atmosphere varies with position and time. The highest values of water vapour pressure (about 30 mb) occur in the tropics near the sea surface.

The lowest values at the earth's surface occur in the high interior of the Antarctic continent during the winter months, when the air is very cold.

On the average, the water vapour pressure decreases with altitude. However, an increase with altitude may sometimes occur in certain parts of the atmosphere.

6.5 The saturation vapour pressure of moist air

Suppose we consider a plane (flat) surface of water at any given temperature. Some of the faster moving molecules of the liquid are usually escaping from the surface of the water into the air above, by the process of evaporation. Some of these return to the water, while others move as a gas in the space above the water.

Ultimately a stage is reached, in which the number of molecules returning to the water in each second will be equal to the number escaping. The space immediately above the water is then said to be saturated at the temperature prevailing in that space.

The partial vapour pressure exerted by the water vapour when the space is saturated is called the saturation vapour pressure at the prevailing temperature.

The saturation vapour pressure varies with the temperature. If the air becomes warmer, more molecules of water vapour are needed to saturate the space above the liquid. Consequently, the partial pressure exerted by the water vapour is greater. Therefore the saturation vapour pressure increases with the temperature.

Warm tropical air has a greater capacity for water vapour than has cold polar air. Hence, high saturation vapour pressures occur in the vicinity of tropical oceans, lakes and rivers.

Table 6.1 indicates some typical saturation vapour pressures over plane water surfaces at various temperatures.

Table 6.1 Saturation vapour pressures over a plane water surface

Temperature (°C)	Saturation vapour pressure (mb)
0	6.11
10	12.27
20	23.37
30	42.43
40	73.77

6.6 The condensation process

If more water vapour is allowed to enter a space which has already been saturated at a given temperature, condensation of the water vapour will occur.

Other factors affect condensation in the atmosphere. Some of these are related to the fact that condensation takes place on small particles in the atmosphere. These are known as condensation nuclei and include dust, smoke, sea salt, ions, etc.

Condensation on some condensation nuclei occurs at vapour pressures below the saturation vapour pressure for a plane water surface at the same temperature. Some condensation nuclei, such as sea salt, also have a strong affinity for water and this fosters condensation. They are called hygroscopic nuclei.

Condensation normally takes place in the atmosphere as the result of the cooling of moist air, i.e. air which contains water vapour. As the temperature becomes lower, less water vapour is needed for the air to become saturated. Eventually a temperature is reached at which the actual vapour pressure becomes equal to the saturation vapour pressure. Any further cooling results in condensation.

6.7 Isobaric processes

The relationships between pressure, temperature and volume in gases were determined by early experiments in physics. In some experiments the pressure of the gas was kept constant.

A physical process in which the pressure of a gas remains constant is called an isobaric process. The word isobaric means equal pressure.

It is also possible to study isobaric processes which occur in moist air. In these processes, a sample of moist air is cooled or warmed at constant pressure. No water vapour is permitted to enter or leave the sample of air. In addition, it is assumed that the water vapour remains in the gaseous state.

If air is cooled isobarically (i.e. at constant pressure), a temperature is reached at which it becomes saturated. This is known as the dew point temperature or simply the dew point.

The dew point temperature may be defined as the temperature to which a sample of moist air has to be cooled at constant pressure for it to become saturated. If the temperature of the air is cooled below the dew point, condensation will occur.

6.8 Adiabatic processes

The cooling for condensation, however, frequently occurs as the result of a change of pressure of the moist air. It is then no longer an isobaric process.

We often study processes in which the pressure, temperature and volume of a gas can all vary, but no heat is added to or removed from the gas. These are adiabatic processes.

Adiabatic processes occur when air moves into regions of lower or higher pressure. In particular, adiabatic cooling occurs when moist air rises. Eventually, the air may become saturated, and any further cooling leads to the condensation of the water vapour and the formation of cloud.

6.9 The freezing process

Pure liquid water, if undisturbed, can be cooled to temperatures well below its freezing point (0°C) and still remain liquid. It is then said to be supercooled.

The introduction of an ice crystal or a freezing nucleus will cause freezing to take place in supercooled water. Freezing will also occur if the supercooled droplets are disturbed.

Laboratory experiments indicate that the lowest temperature at which supercooled droplets can exist is about −40°C. Below this temperature water freezes, regardless of nuclei.

6.10 The deposition process

The process by which water vapour changes directly to ice without passing through the liquid state is called deposition. Sometimes this effect is described as sublimation, but this word is also used to describe the reverse change, i.e. from solid to vapour.

Deposition is not as common as condensation. Nuclei on which deposition takes place are less numerous than condensation nuclei. They are called sublimation nuclei.

It is found that the saturation vapour pressure over ice is slightly less than that over supercooled water at the same temperature. This is indicated in Table 6.2 for plane surfaces.

Table 6.2 Saturation vapour pressure over plane surfaces of water and ice

	Saturation vapour pressure (mb) at a temperature (°C) of				
	−40	−30	−20	−10	0
Over ice	0.128	0.380	1.032	2.597	6.106
Over water	0.189	0.509	1.254	2.862	6.107

The frost point temperature is the temperature to which a sample of moist air has to be cooled at constant pressure for it to become saturated with respect to a plane ice surface. If moist air is cooled below the frost point, water vapour may be deposited as ice on certain solid objects, including other ice surfaces (i.e. on sublimation nuclei).

Notice from Table 6.2 that it might seem possible for this to occur while the air is still unsaturated with respect to water at that temperature. In the atmosphere, however, cloud particles of ice do not usually appear until saturation with respect to water has taken place. It is assumed that condensation first takes place on a certain type of condensation nucleus. If this is of a type which can also act as a freezing nucleus, the water which has condensed on it may freeze. There is still some doubt as to how these nuclei cause freezing, but an essential property of the nucleus seems to be that the structure of the resulting film of water resembles that of an ice crystal.

On the other hand, ice crystals may sometimes be already present, having perhaps fallen from ice clouds at higher levels. If the saturation pressure exceeds the saturation value with respect to ice, direct deposition then takes place on the ice crystals.

6.11 Latent heat

Heat must be added to or subtracted from a substance in order to change its state. While the change of state is taking place, the added heat does not alter the temperature. This energy is stored in the form of latent heat.

For instance, at a pressure of one standard atmosphere water begins to boil at 100°C. As it changes from liquid to gas, the temperature does not rise, even though heat is being added. This heat is needed to separate the molecules and is known as latent heat of vaporisation.

The latent or 'hidden' heat is released again when the water vapour condenses back to the liquid state. In a similar way, latent heat of fusion is required to melt ice, and this is released later if the water freezes to become ice again.

6.12 Moisture indicators

We use the word humidity to describe any measure of the quantity of water vapour contained in a given portion of the atmosphere. For instance, it is sometimes expressed directly as the mass of water vapour in unit volume of air. On the other hand, it may be described indirectly by the contribution which the water vapour makes to the total pressure of all the atmospheric gases.

One important humidity or moisture indicator is the relative humidity.

6.13 Relative humidity

Relative humidity is a convenient moisture indicator. It is the ratio of the mass of water vapour actually present in unit volume of the air to that required to saturate it at the same temperature. Relative humidity is usually expressed as a percentage.

In unsaturated vapours, the mass of water present per unit volume is approximately proportional to the vapour pressure. Hence, the approximate relative humidity may be determined from the relationship:

$$\text{Relative humidity } (\%) = \frac{\text{Actual vapour pressure}}{\text{Saturation vapour pressure at air temperature}} \times 100$$

Many organic substances respond to changes in the relative humidity of the air. Human hair is such a substance, and instruments have been designed to make use of this effect to measure relative humidity.

Notice that the relative humidity may change, even if the water vapour content remains constant. This occurs if the temperature of the sample of air alters.

Under these circumstances, the relative humidity tends to be highest at about dawn, when the temperature is at a minimum. In some situations, saturation may be reached in the air. If condensation occurs, a mist or fog may form. As the temperature rises later in the day the relative humidity decreases, and the mist or fog disappears.

6.14 Methods of measuring humidity

Instruments used for measuring humidity or water vapour content are called hygrometers. Let us consider the various ways in which the humidity of the air at a given locality may be determined.

The dimensions of many organic substances change as the relative humidity varies. This effect is utilised in some hygrometers. For instance, changes in the length of human hair take place as the relative humidity changes. These changes may be magnified by a lever system and used to operate a pointer. This is the principle of the hair hygrometer.

A continuous record of the relative humidity may be obtained if the pointer is replaced by a pen, which marks the reading on a chart attached to a rotating cylindrical drum. This instrument is known as a hair hygrograph.

A simple but accurate method of measuring humidity is to use a psychrometer. This consists merely of two thermometers mounted side by side, one of which measures the air temperature and the other the wet-bulb temperature.

A psychrometer is sometimes called a wet and dry bulb hygrometer. The wet-bulb thermometer is identical with the ordinary dry-bulb thermometer used for measuring air temperature. However, the bulb is covered by a single layer of thin muslin (cotton cloth), which is kept moist by means of a wick (a few strands of thick cotton) dipping into a small vessel of distilled water.

The wet-bulb and dry-bulb thermometers should be ventilated and protected from solar radiation. Psychrometers may be subdivided into stationary screen types and portable types. The latter include Assman and whirling psychrometers.

Once the dry-bulb and wet-bulb temperatures are obtained, tables can be used to determine the relative humidity or the dew point temperature. Let us now consider why this is so.

6.15 The principle of the wet-bulb thermometer

If the air passing the wet-bulb thermometer is not saturated, evaporation of water from the muslin takes place. As the faster moving water molecules escape into the air, they take heat energy with them.

The temperature of the water which remains on the muslin therefore falls. The latent heat of vaporisation is carried away by the molecules that have been converted to water vapour. If evaporation occurs, the wet-bulb temperature will therefore be lower than the dry-bulb reading.

The difference between the temperatures of the dry-bulb and wet-bulb thermometers depends on the rate of evaporation. It is known as the wet-bulb depression. This, in turn, depends on both the relative humidity of the surrounding air and the ventilation around the muslin.

The effect of air motion on evaporation needs to be taken into consideration. The method of ventilation of the psychrometer is therefore important.

6.16 Types of psychrometers

In the simple psychrometer without artificial ventilation the dry-bulb and wet-bulb thermometers are supported vertically in a thermometer screen. The tables used for determining the humidity are usually calculated on the assumption that the average wind speed past the thermometer bulbs is about 1.0 to 1.5 m/s. In practice, the air speeds past the bulb often differ from the assumed range of values. Errors are likely to be greatest when the air is hot and dry, or when the wind is very light.

Several types of artificially ventilated psychrometers are available. The Assman psychrometer requires no screen, the thermometers being protected from radiation by polished metal shields. A stream of air is drawn over the bulbs of the thermometers by a fan.

A fan may also be used to provide ventilation in the aspirated screen type of psychrometer. Air should be drawn past the bulbs at a rate not less than 2.5 m/s and not greater than 10 m/s, if the thermometers are of the types ordinarily used at meteorological stations.

In the whirling psychrometer the two thermometers are mounted side by side in a metal frame. This has a revolving handle which enables it to be whirled. In order to obtain the required ventilation, the bulbs are often insufficiently shielded against radiation. This type of psychrometer should therefore preferably be used in a place shielded from direct solar radiation.

6.17 **Psychrometric tables**

In order to determine the dew point temperature or the relative humidity, it is essential to use the particular psychrometric table which is appropriate to the ventilation speed. Most psychrometric tables are computed on the basis of a ventilation speed from 1 to 10 m/s.

6.18 **Density of moist air**

The density of dry air varies with the pressure and temperature. Near the earth's surface at a pressure of one standard atmosphere (p = 1013.250 mb) and at a temperature of 15°C (T = 288.15 K), the density of dry air is 1.225 kg/m^3.

As indicated by Table 1.1, dry air is a mixture of gases. There is therefore no such thing as a molecule of dry air. However, it is possible to determine the mean molecular weight of dry air. This is sometimes called the 'apparent molecular weight of dry air'. In the homosphere, where the gases are well mixed, it has a value of about 28.96.

Water vapour, on the other hand, has a molecular weight approximately equal to 18. This is only about $\frac{5}{8}$ of the mean molecular weight of dry air in the region up to the mesopause, where most of the gases are well mixed and occur in about the same proportions. A water molecule therefore has less mass than the 'average molecule' of dry air.

Suppose now we take a given volume of dry air and replace some of the molecules by an equivalent number of water vapour molecules. The mass will then be reduced. Since the density is the mass per unit volume, it also will be reduced. Hence, the density of moist air is lower than that of dry air at the same pressure and temperature.

We have seen that the motion of the air is an important factor to be considered when measuring the humidity of the atmosphere. It is also a direct influence on our daily lives, particularly the motion of the air near the earth's suface. In the next chapter we shall investigate the surface wind and the methods used for its measurement.

CHAPTER 7

THE SURFACE WIND

Some of the solar radiation reaching the earth is ultimately transformed into kinetic energy of the gases of the atmosphere. As a result, their molecules are in constant motion.

Wind is air in natural motion. In meteorology, the word usually refers to a rather broad flow of air, either near the earth's surface or in the free atmosphere. In this chapter we shall be concerned with the horizontal flow of the air over the earth's surface.

The flow of air is seldom smooth. Usually turbulence occurs, with eddies of various shapes and sizes developing within the air and interfering with its direct progress. The effect of turbulence near the earth is to produce rapid, irregular changes in both the speed and direction of the wind. These fluctuations occur independently at short intervals of time and constitute the gustiness of the wind.

In this chapter we shall study how the surface wind is measured and consider some of its more important characteristics.

7.1 General principles of surface wind measurement

Wind velocity is a vector quantity, having both a magnitude and a direction. The magnitude of the wind velocity is called the wind speed. The wind direction is regarded as the direction from which it is blowing.

The surface wind velocity is usually subject to rapid fluctuations. The extent to which this occurs is referred to as gustiness.

The surface wind speed, direction and gustiness are best determined from instruments. When this is not practicable, they can be obtained by estimation. It is also desirable to estimate the wind direction when the wind speed is less than two knots, as the instruments are less sensitive to direction changes at low speeds. If there is no appreciable motion of the air, the conditions are said to be calm.

7.2 The exposure of surface wind equipment

It is sometimes difficult to obtain truly representative values of the surface wind speed and direction. The motion of the air is affected by such factors as the roughness of the ground, the type of surface, heat sources, the presence of buildings, etc.

In addition, the wind speed normally increases with height above the earth's surface. It is therefore necessary to specify a standard height for making surface wind measurements, so that winds at different locations can be compared.

The standard exposure of surface wind instruments over level, open terrain is 10 m above the ground. Open terrain is defined as an area where the distance between the instrument and any obstruction is at least ten times the height of the obstruction.

The adoption of the standard exposure is especially important at airports. Where a standard exposure is not available, the surface wind instrument should be installed at such a height that its indications are reasonably unaffected by local obstructions. It should also represent, as far as possible, what the wind at a height of 10 m would be, if there were no obstructions in the vicinity.

7.3 Surface wind direction—units of measurement

The wind direction is regarded as the direction from which the wind is blowing. It is expressed in degrees measured clockwise from geographical north, or in terms of the points of the compass.

For coded reports, however, the wind direction should be expressed on the scale 00-36. Table 7.1 shows the code figures and the exact equivalents in degrees corresponding to the 32 points of the compass.

Table 7.1 Wind direction—compass point equivalents

Compass direction	Exact equivalent in degrees	Code figure	Compass direction	Exact equivalent in degrees	Code figure
Calm	—	00	S by W	191.25	19
N by E	11.25	01	SSW	202.5	20
NNE	22.5	02	SW by S	213.75	21
NE by N	33.75	03	SW	225	23
NE	45	05	SW by W	236.25	24
NE by E	56.25	06	WSW	247.5	25
ENE	67.5	07	W by S	258.75	26
E by N	78.75	08	W	270	27
E	90	09	W by N	281.25	28
E by S	101.25	10	WNW	292.5	29
ESE	112.5	11	NW by W	303.75	30
SE by E	123.75	12	NW	315	32
SE	135	14	NW by N	326.25	33
SE by S	146.25	15	NNW	337.5	34
SSE	157.5	16	N by W	348.75	35
S by E	168.75	17	N	360	36
S	180	18	Variable	—	99

7.4 The measurement of surface wind direction

The surface wind direction is usually indicated by a wind vane. In order to function properly, the vane should be mounted on bearings that will reduce friction to a minimum. It should also be properly balanced about its axis.

Special care should be taken to ensure that the axis of the wind vane is exactly vertical. It should also be correctly orientated with respect to true north.

For synoptic weather reports, the mean direction over a 10-minute period prior to the observation is required. A wind direction recorder is suitable for this purpose.

For aviation purposes, remote recording is desirable. The instruments should be capable of responding to rapid fluctuations in direction. Electrical transmission methods are usually used for this purpose.

It is sometimes necessary to estimate the wind direction. Most wind vanes do not respond to wind direction when the wind speed is less than two knots. In this case, or in the absence of equipment, the direction should be estimated.

7.5 Surface wind speed—units of measurement

The speed of the wind should be given in knots. A knot is equal to a speed of one nautical mile per hour or about 0.51 m/s.

The surface wind speed is rarely constant over any appreciable period of time and it is usually varying rapidly and continuously. The gustiness of the wind produces variations, which are irregular both in period and in amplitude.

For most purposes an average wind is required. For instance, the mean speed of the surface wind over a period of ten minutes is needed for synoptic reports. It should be determined to the nearest knot.

Calm should be reported when the wind speed is less than one knot.

Surface wind speed can be measured in a large variety of ways. The simplest is direct observation of the effect of the wind at the earth's surface without recourse to instruments. The Beaufort wind scale is of this type and was developed by Admiral Sir Francis Beaufort in 1805 for use at sea. Later, it was adapted for use over land. A further development of the scale was to add wind speed equivalents for the various effects.

Wind velocity indicators and recorders have now largely reduced the need for using the Beaufort scale, especially at land stations. Nevertheless, it forms a useful method for estimating wind speeds when other aids are not available. Table 7.2 shows the wind speed equivalents of the various Beaufort numbers.

Table 7.2 Wind speed equivalents

Beaufort number	Description	Wind speed equivalent at a standard height of 10 metres above open flat ground				Specifications for estimating speed over land
		knots (kn)	metres per second (m/s)	kilometres per hour (km/h)	miles per hour (mph)	
0	Calm	<1	0– 0.2	<1	<1	Calm, smoke rises vertically.
1	Light air	1– 3	0.3– 1.5	1– 5	1– 3	Direction of wind shown by smoke drift, but not by wind vanes.
2	Light breeze	4– 6	1.6– 3.3	6– 11	4– 7	Wind felt on face; leaves rustle; ordinary vane moved by wind.
3	Gentle breeze	7–10	3.4– 5.4	12– 19	8–12	Leaves and small twigs in constant motion; wind extends light flag.
4	Moderate breeze	11–16	5.5– 7.9	20– 28	13–18	Raises dust and loose paper; small branches moved.
5	Fresh breeze	17–21	8.0–10.7	29– 38	19–24	Small trees in leaf begin to sway; crested wavelets form on inland waters.
6	Strong breeze	22–27	10.8–13.8	39– 49	25–31	Large branches in motion; whistling heard in telegraph wires; umbrellas used with difficulty.
7	Near gale	28–33	13.9–17.1	50– 61	32–38	Whole trees in motion; inconvenience felt when walking against wind.
8	Gale	34–40	17.2–20.7	62– 74	39–46	Breaks twigs off trees; generally impedes progress.
9	Strong gale	41–47	20.8–24.4	75– 88	47–54	Slight structural damage occurs (chimney-pots and slates removed).
10	Storm	48–55	24.5–28.4	89–102	55–63	Seldom experienced inland; trees uprooted; considerable structural damage occurs.
11	Violent storm	56–63	28.5–32.6	103–117	64–72	Very rarely experienced; accompanied by widespread damage.
12	Hurricane	64 and over	32.7 and over	118 and over	73 and over	Severe and extensive damage.

7.6 The measurement of surface wind speed

Instruments used for measuring the surface wind speed are called anemometers.

The instruments usually used for the measurement of wind speed are of two main types:

(a) Rotation anemometers
(b) Pressure tube anemometers.

The most common wind-speed instrument of the rotation type is the cup anemometer. Three or more specially shaped cups are mounted symmetrically at right angles to a vertical axis. The speed of rotation is dependent on the speed of the wind, irrespective of the direction from which the wind comes.

The cup assembly drives a counting mechanism, and the speed of the wind is computed by using a counting device. A more convenient system is to make the rotor drive a small electric generator. The current is then indicated or recorded by a meter, the scale of which is calibrated in terms of wind speed.

The propellor anemometer is also a rotation type. A propellor is kept facing the wind by a vane and the rotation of the propellor is transmitted to an indicator.

Another rotation type is the windmill anemometer. This drives a counter and therefore a timing device is needed to determine the speed. Windmill anemometers are small and are generally held in the hand. They are mainly used for measuring low wind speeds, such as occur in air conditioning.

The principle of the pressure tube type of anemometer is as follows. A wind vane at the top of the mast keeps the open end of a tube pointing into the wind. The wind blowing into this open end produces an increase in pressure inside it, which depends on the wind speed. This is transmitted to the indicating apparatus by a pressure tube.

An outer tube immediately below the vane is perforated with many small holes. The wind blowing past the holes reduces the pressure inside to an extent depending on the wind speed. This effect is transmitted to the indicator by a suction tube.

The combination of these two effects forms a system, which is independent of any slight pressure differences between the inside and outside of the building in which the recorder is kept.

Two methods of indication are available for pressure tube anemometers. In the Dines float type the pressure and suction are transmitted to a float chamber, where they are made to control the level of a cylinder floating in water. This float, in turn, controls the indicator.

The other method of indication is the aneroid type. This is more suitable for ships where the float type is, of course, useless.

It is also possible to record the wind speed on a chart whose movement is controlled by a clock mechanism. Both the rotation and pressure tube types of anemometers can be converted for use in this way. A self-recording anemometer which makes a continuous record of wind speed measurements is called an anemograph.

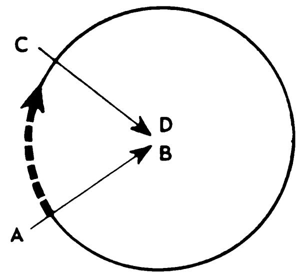
7.7 Variations in the surface wind

Meteorologists use specific terms to describe changes in the wind direction and speed.

When the wind direction changes so that the observer facing the wind has to move in a clockwise direction (i.e. to his right) in order to continue facing the wind, the wind is said to veer. For instance, in Figure 7.1 the wind changes from direction AB to direction CD. The dotted arrow indicates the veering of the wind.

By contrast, if the observer needs to move in an anticlockwise direction, the wind is said to back. Backing of the wind would occur if it changed from direction CD to direction AB.

Figure 7.1 Veering of the wind



When discussing variations in the wind speed, it is necessary to distinguish between a gust and a squall. A gust is a rapid increase in the strength of the wind relative to the mean value obtaining at that time. It is of shorter duration than a squall, and is followed by a lull or slackening of the wind speed.

A squall is a strong wind that commences suddenly, last for some minutes, and then dies away comparatively suddenly. It is more specifically defined as a sudden increase in wind speed by at least 16 knots, the speed increasing to 22 knots or more and lasting for at least one minute.

7.8 Diurnal variation of the surface wind

Places near hills in valleys and close to the sea often experience marked variations in the direction and speed of the surface wind, either in day-time or in night-time. These are called the diurnal (or daily) variations in the surface wind. Some of these local wind systems were discussed in Section 4.10. Others will be mentioned in later chapters.

If the nature of the surface is fairly uniform in an inland area, a marked change in wind speed frequently occurs during the day. It reaches a maximum from about midday until late in the afternoon, due to the transfer of momentum from higher levels down to the surface by convection currents. When the temperature falls late in the afternoon, convection is reduced. The wind speed then decreases and reaches a minimum about dawn.

7.9 Forces acting on moving air

In Section 5.14 we noted that there is normally a difference between the mean sea level pressures at two localities. The pressure gradient is the change in pressure with horizontal distance measured in the direction from high to low pressure.

One might expect that the pressure gradient would lead to the movement of a parcel of air in the direction from high to low pressure. Actually, a pressure gradient force does exist in this direction, but the air is seldom observed to move along this path. Why is this so?

A number of factors may be involved and one of these is the rotation of the earth. To an observer on the earth's surface the moving air will appear to be deflected. This is the Coriolis effect, and arises from the fact that an observer, standing on the earth's surface, rotates at the same rate as the earth spins on its axis.

The apparent force acting on a parcel of air is known as the Coriolis force. Its value is zero at the equator, but increases with latitude to its maximum value at the poles. Its magnitude is also proportional to the speed of the air relative to the earth's surface.

In the northern hemisphere, the Coriolis force appears to deflect the air towards the right, when viewed along the direction of motion; in the southern hemisphere, the deflection is to the left.

Another force acting on the parcel of air is the friction force. This force arises whenever relative motion occurs between the air and the earth's surface, or between adjacent layers of air. It acts in the direction opposite to that of the relative movement.

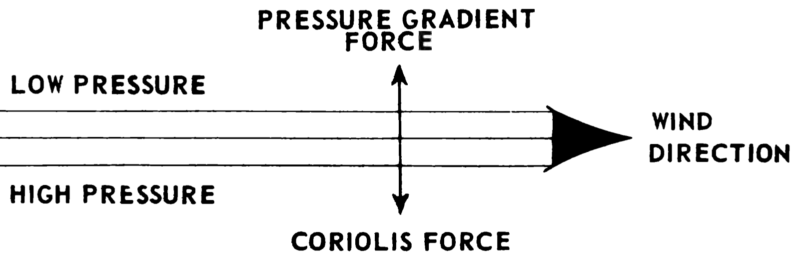
The friction force is largest near the earth's surface, and air below a height of about 1 km is said to lie in the friction layer. Above the friction layer, the friction force is often negligible, and this region is called the free atmosphere.

7.10 **The geostrophic wind**

Let us consider a parcel of air moving horizontally at a constant speed in a region where the friction force is negligible. The horizontal forces acting on it are the pressure gradient force and the Coriolis force.

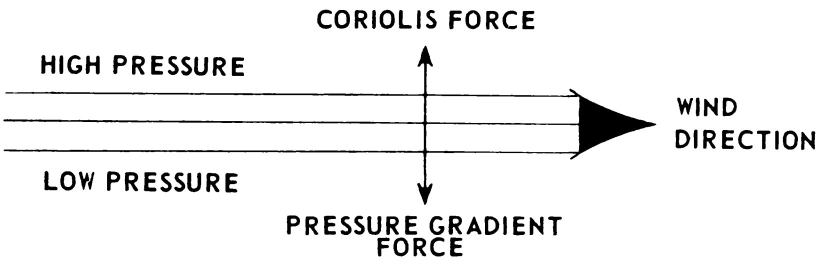
If the magnitudes of these two forces are exactly equal, the pressure gradient force will prevent the Coriolis force from causing a deflection either to the left or to the right. Motion of this type is known as geostrophic flow. Figure 7.2 shows how the forces may balance in the northern hemisphere.

Figure 7.2 Geostrophic flow in the northern hemisphere



The corresponding situation in the southern hemisphere is shown in Figure 7.3.

Figure 7.3 Geostrophic flow in the southern hemisphere



Because geostrophic flow is horizontal, the air actually flows along a great circle. When the motion is projected onto the flat surface of a synoptic weather chart, it appears to be in a straight line.

For a given latitude, there is a certain wind speed at which the Coriolis force just balances the existing pressure gradient force. This is called the geostrophic wind speed, and its value may be determined, at any point on a synoptic chart where the distance between adjacent isobars is known, using a geostrophic wind scale.

The geostrophic wind speed cannot be determined at the equator, where the Coriolis force is zero. If no other forces were acting, the air would then move in the direction of the pressure gradient force, i.e. from high to low pressure.

Actually, the Coriolis force is small in low latitudes and geostrophic flow seldom occurs between latitudes 15°N and 15°S.

7.11 The gradient wind

Let us again consider the motion of a parcel of air moving horizontally at constant speed in a region where the friction force is negligible.

Suppose that in this case the magnitudes of the pressure gradient force and the Coriolis force are not equal. The motion will then be curved either to the left or to the right.

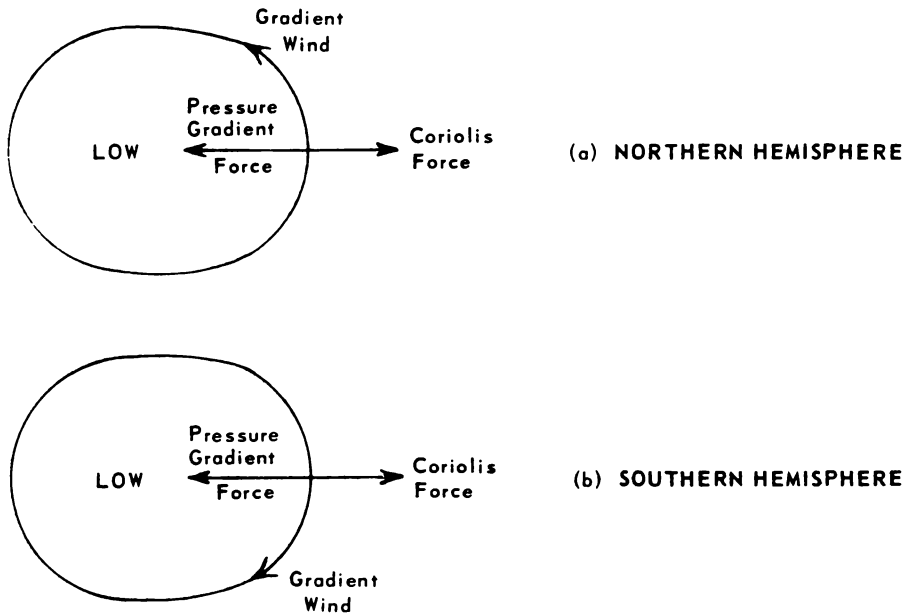
At a given latitude, the Coriolis force may be either larger or smaller than the pressure gradient force. This will depend on the wind speed.

Frictionless, horizontal motion at constant speed is called gradient flow. This implies that the flow is tangential to the isobar at each point. The wind speed at which this occurs for a given latitude and pressure gradient is called the gradient wind speed.

The flow may be gradient merely at a single point or else over an extended path. In the former case, its direction is tangential to the isobar only at the given point. For extended flow, the motion is along the isobars throughout the entire trajectory, but this can only occur if the isobars are not changing with time. Its direction is then tangential to the isobar at each point along its length.

If gradient flow occurs over an extended path, the air will move at constant speed in the direction of the curved isobars of a synoptic chart. If the pressure gradient force is greater than the Coriolis force, the flow will be curved around a low pressure area. Diagram (a) of Figure 7.4 shows how this may occur in the northern hemisphere. The corresponding situation in the southern hemisphere is shown in diagram (b).

Figure 7.4 Gradient flow around a low pressure area



Notice that the flow is anticlockwise in the northern hemisphere, because the Coriolis force acts to the right. By contrast, the motion in the southern hemisphere around a low pressure area is clockwise. In both cases, however, the motion is called cyclonic flow.

Cyclonic flow is a flow in the direction of the earth's rotation. If a globe of the earth spins in the direction of the earth's rotation the motion appears to be anti-clockwise when viewed from above the north pole. By contrast, the rotation seems to be clockwise when viewed from above the south pole.

You should now draw diagrams to show that if the pressure gradient force is less than the Coriolis force, the motion is curved around the region where the pressure is higher. It will now be clockwise in the northern hemisphere and anticlockwise in the southern hemisphere.

The flow is therefore in the opposite direction to the earth's rotation in each hemisphere. Flow around a high pressure area is therefore called anticyclonic flow.

As in the case of geostrophic flow, friction has been neglected. However, isobars are normally curved. Hence, gradient flow is usually a closer approximation to the actual motion of the free atmosphere than is geostrophic flow.

The magnitude of the gradient wind at any latitude depends not only on the pressure gradient, but also on the radius of curvature of the isobars. This is frequently difficult to measure accurately.

7.12 **Buys Ballot's law**

In 1857 a Dutch meteorologist, Buys Ballot, described a relationship which exists between the wind direction and the isobars. If you refer to Figures 7.2 and 7.4 (a) you will be able to determine this relationship for the northern hemisphere. On standing with your back to the wind, the lower pressure is on your left.

In the southern hemisphere, the lower pressure is always on the right. This can be checked by studying Figures 7.3 and 7.4 (b).

This relationship is known as Buys Ballot's law:

If an observer stands with his back to the wind the lower pressure is on his left in the northern hemisphere, and on his right in the southern hemisphere.

7.13 **Wind in the friction layer**

The surface wind is normally measured at the standard level of 10 m above the earth's surface. In practice, this wind differs appreciably from geostrophic wind speed computed from a mean sea level isobaric chart, even for straight isobars.

This is due to the fact that the friction force cannot be neglected in the friction layer. As a result, the observed wind speed is on the average only about one-third of the geostrophic speed over land and about two-thirds over the sea.

The direction of the surface wind is not along the isobars, due to the action of the friction force. The lower wind speed reduces the Coriolis force, and the pressure gradient force causes the air near the earth's surface to flow somewhat across the isobars from high to low pressure. The friction force may lead to a cross-isobar flow at an angle of about 30° over land and about 10° over the sea.

The effect of the friction force is greatest at the ground, where wind flows over a rough surface and produces turbulence. It decreases rapidly upwards until it becomes negligible at the top of the friction layer. This occurs at a height of approximately 1 km above the ground.

7.14 **Horizontal mass convergence and divergence**

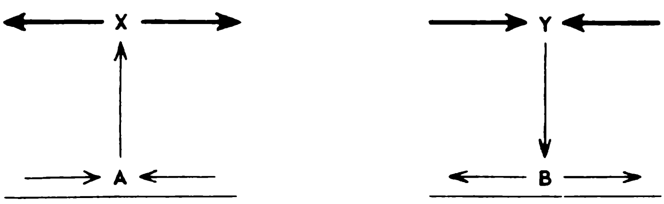
If there is a net gain or loss of mass of air above a given horizontal area at the earth's surface, the pressure will change at the bottom of the column. A net loss of mass from the column leads to a fall in pressure at the earth's surface; a net gain of mass is associated with a pressure rise.

When there is a net horizontal flow of air into a region, horizontal (mass) convergence is said to occur. If convergence were to continue there would be an accumulation of air in the region and the density of the air would rise. However, observed density changes are small because vertical motion tends to take place. In this way, an accumulation of mass is prevented from occurring.

Horizontal (mass) divergence is the opposite process. Divergence takes place when there is a net horizontal outflow from the region considered. It also leads to vertical motion of the atmosphere.

Developing depressions and troughs of low pressure are associated with convergence in the lower troposphere and divergence aloft. Of course, the effect of divergence aloft must be greater than that of convergence near the earth's surface if the surface pressure is to fall. Figure 7.5 shows how horizontal (mass) divergence at

Figure 7.5 Horizontal mass convergence and divergence associated with vertical motion



x is associated with horizontal (mass) convergence at A. This is linked with upward motion.

By contrast, developing anticyclones and ridges of high pressure are associated with convergence aloft and divergence in the lower troposphere. These effects are linked with downward motion of the air. Figure 7.5 shows horizontal (mass) convergence at y is greater than the horizontal (mass) divergence at B, if the surface pressure is increasing.

If a rapid fall of pressure occurs at A, a low pressure area develops and the air is accelerated towards that region by the action of the pressure gradient force. The Coriolis force, being proportional to the wind speed, is at first small, and so a balanced flow is not immediately achieved. Slow upward motion (ascent) may then take place over a wide area. Thick masses of cloud and precipitation may occur if the moisture content of the air is high.

By contrast, a rapid rise of pressure near B leads to the development of a high pressure area in that region. Slow downward motion (subsidence) over a wide area inhibits cloud development.

Even when the Coriolis force has increased, surface friction still leads to some cross-isobar flow from high to low pressure. Low level convergence is associated with low pressure centres and divergence with anticyclones. These effects are, however, confined to the friction layer and the resulting vertical motion is relatively small. While surface friction may account for the formation of a certain amount of cloud in depressions, it is entirely inadequate to explain any but the lightest rainfall.

7.15 The advection of air

The atmosphere is a medium in which mass motions are easily started. In this way, heat may be transferred either by horizontal motion or by vertical movement of the air.

In meteorology, the term convection is often used to indicate vertical motion. The magnitude of the vertical velocities, however, are usually only about 1% of the horizontal wind speeds.

Horizontal motion generally occurs on a much greater scale. It may even involve the transfer of heat energy from tropical to polar regions over distances of thousands of kilometres.

The transfer of heat and other physical quantities horizontally by the wind is known as advection. The word has a Latin derivation and means 'carry towards'. Advection currents are usually greater and more sustained than vertical currents.

Vertical motion, however, has important effects. For instance, it may lead to the development of clouds in the atmosphere. In the next chapter we shall study how clouds can be distinguished from one another.

CHAPTER 8

CLOUD CLASSIFICATION

Clouds are continuously in the process of evolving in the atmosphere. When you view the skies, you will notice that they occur in an unlimited number of forms.

As the water vapour in the atmosphere changes its state, water droplets and ice particles are formed, and these become visible to our eyes. Sometimes we see white clouds high in the sky; sometimes low grey or dark clouds cover large parts of the heavens.

Our weather depends greatly on the types of clouds which form in the skies. For this reason, meteorologists direct a great deal of attention towards the study of the structure and development of clouds.

Recently, they have been assisted in their studies by the great advances which have occurred as a result of the launching of earth satellites. Cloud photographs are now available for large regions of the earth. These supplement the information which we receive from aircraft, and so our knowledge of the clouds has been increased considerably.

In this chapter we shall investigate a limited number of cloud forms, so that we may discuss some of the more important characteristics of the weather. Later, we shall consider cloud types in more detail when we study weather observations.

8.1 Cloud names

There are two main types of clouds. Cumuliform clouds comprise cumulus-type formations in which the clouds are usually separated from each other by clear spaces. By contrast, straiform clouds form sheets or layers covering large areas of the sky.

The names of clouds are descriptive of their type and form. A number of Latin words are used to indicate certain characteristics of clouds. The word 'nimbus' (a shower) is added to the names of clouds which produce precipitation. The prefix 'fracto-' (broken) appears at the beginning of the names of broken, wind-blown clouds. The word 'cirrus' (a hair) is used to indicate a cloud of hair-like appearance.

8.2 Cloud genera

Ten main groups of clouds can be distinguished. Each is called a genus, the plural being denoted by the word genera.

Each genus may be further subdivided into species and varieties. However, we shall not consider these subdivisions at this stage.

Clouds are grouped into the ten following genera:

(a) Cirrus
(b) Cirrocumulus
(c) Cirrostratus
(d) Altocumulus

(e) Altostratus
(f) Nimbostratus
(g) Stratocumulus
(h) Stratus
(i) Cumulus
(j) Cumulonimbus.

8.3 Height, altitude and vertical extent

It is often important to refer to the level at which certain parts of a cloud occur. Two concepts are used to identify such a level, namely height and altitude.

The height of a point (for example the base of a cloud) is the vertical distance from the point of observation to the level of that point. Note that the point of observation may sometimes be on a hill or mountain.

The altitude of a point is the vertical distance measured from mean sea level to the level of that point.

Surface observers generally use the concept of height. Observers on aircraft, however, usually refer to altitude.

The vertical extent of a cloud is the vertical distance between the level of its base and that of its top.

8.4 Etages

Most clouds are generally encountered over a range of altitudes varying from sea level to the level of the tropopause. The altitude of the tropopause varies in time and space. The tops of the clouds are therefore higher in the tropics than in middle and high latitudes.

By convention, the part of the atmosphere in which clouds are usually present has been vertically divided into three étages or intervals—high, middle and low.

Each étage is defined by the range of levels at which clouds of certain genera occur most frequently. The étages overlap and their limits vary with latitude. The approximate heights of the limits in kilometres are as follows:

Etages	Polar regions	Temperate regions	Tropical regions	Australian regions
High	3–8 km	5–13 km	6–18 km	Above 6 km
Middle	2–4 km	2– 7 km	2– 8 km	2.5–6 km
Low	From the earth's surface to 2 km	From the earth's surface to 2 km	From the earth's surface to 2 km	From the earth's surface to 2.5 km

The étages in which six of the genera are found are as follows:

(a) Cirrus, cirrocumulus and cirrostratus for the high étage (high level clouds).
(b) Altocumulus for the middle étage (middle level clouds).
(c) Stratocumulus and stratus for the low étage (low level clouds).

With regard to the other four genera, the following remarks may be made:

(a) Altostratus is usually found in the middle étage, but it often extends to higher levels.
(b) Nimbostratus is almost invariably found in the middle étage, but it usually extends both downwards into the low étage and upwards into the upper étage.
(c) Cumulus and cumulonimbus usually have their bases in the low étage, but their vertical extent is often so great that their tops may reach into the middle and high étages.

When the height of a particular cloud is known, the concept of étages may be of some help to the observer in identifying it. Its genus can then be determined, by making a choice from among the genera normally encountered in the étage corresponding to its height.

8.5 **Definitions of the ten cloud genera**

The World Meteorological Organization has set out formal definitions of the ten cloud genera in the *International Cloud Atlas*, Volume I. Typical clouds are illustrated in Figure 8.1.

The definitions and the recognised two-letter abbreviations for the ten cloud genera are as follows:

Cirrus (Ci)
Detached clouds in the form of white, delicate filaments or white or mostly white patches or narrow bands. These clouds have fibrous (hair-like) appearance, or a silky sheen, or both.

Cirrocumulus (Cc)
Thin, white patch, sheet or layer of cloud without shading, composed of very small elements in the form of grains, ripples, etc., merged or separate, and more or less regularly arranged; most of the elements have an apparent width of less than one degree.

Cirrostratus (Cs)
Transparent, whitish cloud veil of fibrous (hair-like) or smooth appearance, totally or partly covering the sky, and generally producing halo phenomena.

Altocumulus (Ac)
White or grey, or both white and grey, patch, sheet or layer of cloud, generally with shading, composed of laminae, rounded masses, rolls, etc., which are sometimes partly fibrous or diffuse and which may or may not be merged; most of the regularly arranged small elements usually have an apparent width of between one and five degrees.

Altostratus (As)
Greyish or bluish cloud sheet or layer of striated, fibrous or uniform appearance, totally or partly covering the sky, and having parts thin enough to reveal the sun at least vaguely, as through ground glass. Altostratus does not show halo phenomena.

Nimbostratus (Ns)
Grey cloud layer, often dark, the appearance of which is rendered diffuse by more or less continuously falling rain or snow, which in most cases reaches the ground. It is thick enough throughout to blot out the sun.

Low, ragged clouds frequently occur below the layer, with which they may or may not merge.

Stratocumulus (Sc)
Grey or whitish, or both grey and whitish, patch, sheet or layer of cloud which almost always has dark parts, composed of tessellations, rounded masses, rolls, etc., which are non-fibrous (except for virga) and which may or may not be merged; most of the

regularly arranged small elements have an apparent width of more than five degrees.

Stratus (St)
Generally grey cloud layer with a fairly uniform base, which may give drizzle, ice prisms or snow grains. When the sun is visible through the cloud, its outline is clearly discernible. Stratus does not produce halo phenomena except, possibly, at very low temperatures.

Sometimes stratus appears in the form of ragged patches.

Cumulus (Cu)
Detached clouds, generally dense and with sharp outlines, developing vertically in the form of rising mounds, domes or towers, of which the bulging upper part often resembles a cauliflower. The sunlit parts of these clouds are mostly brilliant white; their base is relatively dark and nearly horizontal.

Sometimes cumulus is ragged.

Cumulonimbus (Cb)
Heavy and dense cloud, with a considerable vertical extent, in the form of a mountain or huge towers. At least part of its upper portion is usually smooth, or fibrous or striated, and nearly always flattened; this part often spreads out in the shape of an anvil or vast plume.

Under the base of this cloud, which is often very dark, there are frequently low ragged clouds either merged with it or not, and precipitation sometimes in the form of virga.

8.6 **Cloud recognition**

Although the classification of clouds into typical forms is of great use, the problem of identifying cloud forms is not always an easy one. It does not overcome the difficulties that arise from the gradual transition between the various types of clouds.

Natural cloud does not always conform to the types described in the artificial classification mentioned in the previous section. At times, the clouds may be intermediate between two types. In these cases, the experience and judgment of the observer becomes important.

Reliable observations of cloud can best be made, by keeping a close and continuous watch on their development. It is not always sufficient merely to make a brief examination of the sky at the observation hour.

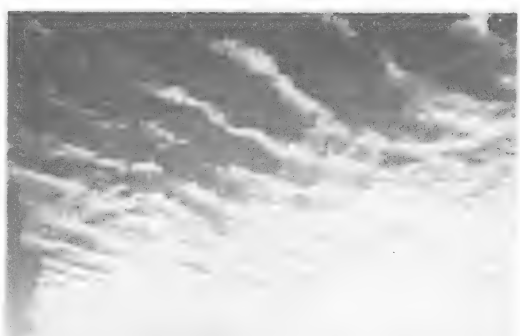
Water droplets and ice crystals occur in the atmosphere in forms other than those of clouds. In the next chapter we shall study the various products of the condensation and deposition of water vapour that are found in the atmosphere.

Figure 8.1 Examples of the ten cloud genera

(a) Cirrus originating in tufts progressively invading the sky.

(b) Cirrocumulus.

(c) Veil of Cirrostratus covering the whole sky, with accompanying halo phenomena.



(d) Altocumulus in the form of waves. Some small globular elements of Altocumulus also present. Patches of Stratocumulus below.

Cloud photographs by G. E. Tralaggan

(e) Altostratus (thin). Sun appears as through ground glass. Fractocumulus below.

(f) Nimbostratus with Fractocumulus below.

(g) Stratocumulus not formed from Cumulus.

(h) Layer of Stratus.

(i) Cumulus of moderate vertical development.

(j) Cumulonimbus with anvil top. Cumulus also present.

CHAPTER 9

HYDROMETEORS

Features of the weather, such as rain, snow and hail, provide ample evidence that water droplets and ice crystals occur in the atmosphere in forms other than those of clouds. Liquid and solid particles may be suspended in the air, or fall towards the earth's surface. Sometimes they may be blown into the atmosphere by the action of the wind. On other occasions, water droplets or ice crystals may be deposited on objects on the ground or in the free air.

In this chapter we shall study the characteristics of liquid and solid particles which are of concern to meteorologists. We shall also consider how they may be distinguished from one another.

9.1 **Definition of a meteor**

Most dictionaries define meteors as visible or optical phenomena in the atmosphere. Astronomical meteors entering the atmosphere from space are so commonly seen as 'shooting stars', that most people think of these objects when the word 'meteor' is mentioned.

However, non-astronomical meteors are even more common. For this reason, meteorologists define the word 'meteor' in a special way, which you should not confuse with the meaning given to it in astronomy.

In meteorology, the term meteor refers to any phenomenon, other than a cloud, observed in the atmosphere or on the surface of the earth. It may consist of a precipitation or a deposit of liquid or solid particles. It also includes optical or electrical phenomena which occur in the atmosphere.

There are many types of meteors. However, by considering the nature of the particles or the physical processes involved in their occurrence, four main types can be distinguished:

(a) hydrometeors
(b) lithometeors
(c) photometeors
(d) electrometeors.

9.2 **Definitions of meteors**

The World Meteorological Organization defines these four types of meteors in the *International Cloud Atlas*, Volume I. These definitions are as follows:

Hydrometeor

A hydrometeor is a meteor consisting of an ensemble of liquid or solid aqueous particles, falling through or suspended in the atmosphere, blown by the wind from the earth's surface, or deposited on objects on the ground or in the free air.

Lithometeor
A lithometeor is a meteor consisting of an ensemble of particles, most of which are solid and non-aqueous. The particles are more or less suspended in the air, or lifted by the wind from the ground.

Photometeor
A photometeor is a luminous phenomenon produced by the reflection, refraction, diffraction or interference of light from the sun or the moon.

Electrometeor
An electrometeor is a visible or audible manifestation of atmospheric electricity.

Lithometeors include dust, smoke, etc., and photometeors occur as rainbows, haloes, etc. Lightning, thunder and thunderstorms are included in the various types of electrometeors.

In this chapter we shall be mainly concerned with hydrometeors. The other meteors will be studied at a later stage.

9.3 **Classification of hydrometeors**

(a) Groups or ensembles of falling particles. These originate mostly in clouds and include drizzle, rain, snow, snow pellets and ice prisms.
(b) Groups or ensembles of falling particles which evaporate or sublime before they reach the earth's surface (virga).
(c) Hydrometeors consisting of ensembles of particles suspended in the air. Fog and mist are similar to clouds, but are considered as meteors because of their occurrence at or near the earth's surface.
(d) Liquid or solid particles blown from the earth's surface by the wind. Drifting snow, blowing snow and spray are usually confined to the lowest layers of the atmosphere.
(e) Liquid or solid particles deposited on exposed objects. These include dew, hoar-frost, rime and glaze. In hoar-frost and rime the individual groups of solid particles can usually be seen, even though they may fuse together. No particle structure can be discerned in glaze and it forms in homogeneous and smooth layers.

9.4 **Precipitation**

Precipitation refers to falling hydrometeors that eventually reach the ground. It does not include a virga, which is seen streaming from the bottom of a cloud, but does not reach the earth's surface.

When precipitation reaches the ground, it may be measured as a certain depth of water. This measurement is described as rainfall, irrespective of the form of the precipitation when it first reached the earth's surface.

The intensity of precipitation is judged by the rate at which the precipitation would accumulate on a horizontal surface, if prevented from running away. It is also the rate at which precipitation would be collected in a rain gauge. It may be described as being light, moderate or heavy, but these words are interpreted in various ways by different persons.

Those hydrometeors which consist of falling particles occur either as showers, or in the form of more or less uniform (intermittent or continuous) precipitation. A distinction should be made between showers and intermittent precipitation.

Showers are characterised by their abrupt beginning and end. There are generally rapid and sometimes violent variations in the intensity of the precipitation. In general, the water drops and solid particles falling in a shower are larger than those falling in non-showery precipitation.

Intermittent precipitation is precipitation which is not continuous at the earth's surface. However, the cloud from which it comes is more or less continuous. It differs from showers in that it does not commence and end suddenly. There is no clearing of the cloud between the onset and cessation of the non-showery precipitation.

Whether the hydrometeors occur as showers or not depends on the clouds in which they originate. Showers fall from dark cumuliform clouds, such as Cumulonimbus; non-showery precipitation usually falls from stratiform clouds, mainly Altostratus and Nimbostratus.

It is therefore often possible to identify clouds by the character of their precipitation. This can be very useful at night and in doubtful cases which occur by day.

9.5 **Definitions and descriptions of hydrometeors**

In order to distinguish the various hydrometeors from one another it is necessary to define them specifically. Their definitions and general characteristics are set out in the *International Cloud Atlas*, Volume I. Details are as follows:

Rain

Precipitation of liquid water particles, either in the form of drops of more than 0.5 mm diameter, or of smaller widely scattered drops.

Raindrops are normally larger than drops of drizzle. Nevertheless, drops falling on the edge of a rain zone may be as small as drizzle drops, owing to partial evaporation; rain is then distinguished from drizzle by the fact that the drops are scattered. In certain cases, clouds may include an abnormally large number of fine particles of dust or sand, lifted from the ground in dust storms or sandstorms. These particles may fall to the ground with the raindrops (mud rain), often after having been carried over great distances.

Freezing rain

Rain, the drops of which freeze on impact with the ground or with objects on the earth's surface or with aircraft in flight.

Drizzle

Fairly uniform precipitation composed exclusively of fine drops of water (diameter less than 0.5 mm) very close to one another.

The drops appear almost to float, thus making visible even slight movements of the air.

Drizzle falls from a fairly continuous and dense layer of Stratus, usually low, sometimes even touching the ground (fog). The amount of precipitation in the form of drizzle is sometimes considerable (up to 1 mm per hour), especially along coasts and in mountainous areas.

Freezing drizzle

Drizzle, the drops of which freeze on impact with the ground or with objects on the earth's surface or with aircraft in flight.

Snow
Precipitation of ice crystals, most of which are branched (sometimes star-shaped).

The branched crystals are sometimes mixed with unbranched crystals. At temperatures higher than about $-5°C$, the crystals are generally agglomerated into snowflakes.

Snow pellets
Precipitation of white and opaque grains of ice. These grains are spherical or sometimes conical; their diameter is about 2.5 mm.

The grains are brittle and easily crushed; when they fall on hard ground they bounce and often break up. Precipitation of snow pellets generally occurs in showers together with precipitation of snowflakes or raindrops, when surface temperatures are around 0°C.

Snow grains
Precipitation of very small white and opaque grains of ice. These grains are fairly flat or elongated; their diameter is generally less than 1 mm.

When the grains hit hard ground they do not bounce or shatter. They usually fall in very small quantities, mostly from Stratus or from fog, and never in the form of a shower.

Ice pellets
Precipitation of transparent or translucent pellets of ice, which are spherical or irregular, rarely conical, and which have a diameter of 5 mm or less.

The pellets of ice usually bounce when hitting hard ground and make a sound on impact; they may be subdivided into two main types:

(a) Frozen raindrops or largely melted and refrozen snowflakes. The freezing process usually takes place near the earth's surface.

(b) Pellets of snow encased in a thin layer of ice which has formed from the freezing either of droplets intercepted by the pellets or of water resulting from the partial melting of the pellets.

Hail
Precipitation of small balls or pieces of ice (hailstones) with a diameter ranging from 5 to 50 mm or sometimes more, falling either separately or agglomerated into irregular lumps.

Hailstones are composed, almost exclusively, of transparent ice, or of a series of transparent layers of ice, at least 1 mm in thickness, alternating with translucent layers. Hail falls are generally observed during heavy thunderstorms.

Ice prisms
A fall of unbranched ice crystals, in the form of needles, columns or plates, often so tiny that they seem to be suspended in the air. These crystals may fall from a cloud or from a cloudless sky.

The crystals are visible mainly when they glitter in the sunshine (diamond dust); they may then produce a luminous pillar or other halo phenomena. This hydrometeor, which is frequent in polar regions, occurs at very low temperatures and in stable air masses.

Fog

A suspension of very small water droplets in the air, reducing the horizontal visibility at the earth's surface to less than 1 km.

When sufficiently illuminated, individual fog droplets are frequently visible to the naked eye; they are then often seen to be moving in a somewhat turbulent manner. The air in fog usually feels raw, clammy or wet. Relative humidity in fog is generally near 100%.

This hydrometeor forms a whitish veil which covers the landscape; when mixed with dust or smoke it may, however, take on a faint coloration, often yellowish. In the latter case, it is generally more persistent than when it consists of water droplets only.

Ice fog

A suspension of numerous minute ice crystals in the air, reducing the visibility at the earth's surface.

The crystals often glitter in the sunshine. Ice fog produces optical phenomena such as luminous pillars, small haloes, etc.

Note: In the International Codes for weather reports, the term 'fog' is used when the hydrometeor fog or mist reduces the horizontal visibility at the earth's surface to less than 1 km.

Mist

A suspension in the air of microscopic water droplets or wet hygroscopic particles, reducing the visibility at the earth's surface.

The air in mist usually does not feel raw or clammy. Relative humidity in mist is generally lower than 100%.

This hydrometeor forms a generally fairly thin, greyish veil which covers the landscape.

Note: In the International Codes for weather reports, the term 'mist' is used when the hydrometeor mist reduces the horizontal visibility at the earth's surface to not less than 1 km.

Drifting snow and blowing snow

An ensemble of snow particles raised from the ground by a sufficiently strong and turbulent wind.

The wind conditions (speed and gustiness) necessary to produce these hydrometeors depend on the state and age of the surface snow.

(a) Drifting snow

An ensemble of snow particles raised by the wind to small heights above the ground. The visibility is not sensibly diminished at eye level.

Very low obstacles are veiled or hidden by the moving snow. The motion of the snow particles is more or less parallel to the ground.

(b) Blowing snow

An ensemble of snow particles raised by the wind to moderate or great heights above the ground. The horizontal visibility at eye level is generally very poor.

The concentration of the snow particles may sometimes be sufficient to veil the sky and even the sun. The snow particles are nearly always violently stirred up by the wind.

Spray
An ensemble of water droplets torn by the wind from the surface of an extensive body of water, generally from the crest of waves, and carried up a short distance into the air.

When the water surface is rough, the droplets may be accompanied by foam.

Dew
A deposit of water drops on objects at or near the ground, produced by the condensation of water vapour from the surrounding clear air.

Dew is formed when:

(a) Exposed surfaces are cooled below the dew point of the ambient air; such cooling is usually due to nocturnal radiation and the dew is deposited mainly on objects at or near the ground.
(b) Warm moist air comes in contact with a colder surface, the temperature of which is lower than the dew point of the air; this generally happens as a result of advection of air.

White dew
A deposit of white frozen dew drops.

Hoar-frost
A deposit of ice having a crystalline appearance, generally assuming the form of scales, needles, feathers or fans.

This hydrometeor is produced in a manner similar to dew, but at a temperature below 0°C.

Rime
A deposit of ice, composed of grains more or less separated by trapped air, sometimes adorned with crystalline branches.

Rime is produced by the rapid freezing of supercooled, very small water droplets; it sometimes grows to thick layers.

Near the ground, rime is deposited on objects, mainly on the sides exposed to the wind and mostly on points and edges, by the freezing of supercooled fog droplets or (in mountainous regions) of supercooled cloud droplets.

In the free atmosphere, rime occurs as a type of icing, friable and resembling caked pellets of snow, on the parts of an aircraft exposed to the relative wind.

Glaze (clear ice)
A generally homogeneous and transparent deposit of ice formed by the freezing of supercooled drizzle droplets or raindrops on objects the surface temperature of which is below or slightly above 0°C.

It may also be produced by the freezing of non-supercooled drizzle droplets or raindrops immediately after impact with surfaces the temperature of which is well below 0°C.

Glaze is observed at the earth's surface when raindrops fall through a layer of subfreezing temperature of sufficient depth.

In the free atmosphere, glaze occurs as a compact and smooth type of icing (clear ice). It is sometimes encountered in clouds containing large supercooled droplets, which freeze after impact with the parts of an aircraft exposed to the relative wind;

on other occasions, glaze may cover all parts of an aircraft which are exposed to supercooled precipitation.

Note: Glaze on the ground must not be confused with 'ground ice', which is formed when: (i) water from a precipitation of non-supercooled drizzle droplets or raindrops later freezes on the ground; (ii) snow on the ground freezes again after having completely or partly melted; (iii) snow on the ground is made compact and hard by traffic.

Spout

A phenomenon consisting of an often violent whirlwind, revealed by the presence of a cloud column or inverted cloud cone (funnel cloud), protruding from the base of a Cumulonimbus, and of a 'bush' composed of water droplets raised from the surface of the sea or of dust, sand or litter, raised from the ground.

The axis of the funnel cloud is vertical inclined or sometimes sinuous. Not uncommonly, the funnel merges with the bush.

The air in the whirlwind rotates rapidly, most often in a cyclonic sense; a rapid rotary movement may also be observed outside the funnel and the 'bush'. Further away, the air is often very calm.

The diameter of the cloud column, which is normally of the order of 10 m, may in certain regions occasionally reach some hundreds of metres.

Several spouts may sometimes be observed connected with one single cloud.

Spouts are often very destructive in North America (tornadoes), where they may leave a path of devastation up to 5 km wide and several hundred kilometres long.

Weak spouts are occasionally observed under Cumulus clouds.

9.6 Precipitation associated with cloud genera

When falling particles reach the place of observation, it is possible to determine their nature. As these hydrometeors are closely associated with certain cloud genera, their identification often enables the observer to determine which clouds are present. Table 9.1 lists the various precipitation particles and the cloud genera in which they originate.

Table 9.1 Precipitation associated with cloud genera

	Cloud genera					
Hydrometeors	As	Ns	Sc	St	Cu†	Cb†
Rain	+	+	+		+	+
Drizzle			+	+		
Snow	+	+				+
Snow pellets						+
Ice pellets						+
Ice prisms*				+		

* An aircraft observer may encounter ice prisms under cirrus, cirrocumulus, cirrostratus and altocumulus clouds.
† Cumulus and cumulonimbus clouds give precipitation in the form of showers.

9.7 The rate of fall of water droplets

An object falling through the air is subjected to three forces:

(a) The force of gravity
(b) The upthrust due to buoyancy (Archimedes' principle)
(c) The air resistance due to its motion.

At first it accelerates, because the force of gravity is greater than the combined upward effect of the other two forces. However, the faster the object travels towards the earth, the greater the air resistance becomes.

Eventually, if the object has not already reached the ground, the forces become balanced and it then descends at a constant velocity. This is known as the terminal velocity.

A parachutist at first accelerates downwards, but eventually descends at a constant speed, soon after his parachute has opened. Falling leaves also reach a terminal velocity. In the same way, falling raindrops accelerate to a maximum downward velocity, and then continue on at the the same speed.

The larger the particle of water, the greater will be its terminal velocity. Table 9.2 shows this effect.

Table 9.2 Terminal velocity of water droplets

Diameter micron (μ)	Terminal velocity m/s
2	0.00012
8	0.00192
100	0.27
200	0.72
500	2.06
1000	4.03
2000	6.49
5000	9.09
5800	9.17

($1\mu = 10^{-3}$ mm $= 10^{-6}$ m)

9.8 The deformation of falling raindrops

Water droplets are deformed as they fall through the air. The amount of deformation increases with the size of the drop.

This accounts for the fact that rainbows are only seen with certain sizes of water drops. Rainbows result from the dispersion of the visible radiation from the sun into the various wavelengths, which comprise the colours of the rainbow.

The dispersion is associated with the refraction and internal reflection of sunlight inside spheres of water. However, if the size of a drop exceeds a certain size, its shape is no longer spherical. It becomes rather flattened underneath and dispersion is not observed.

Even if the raindrops are spherical, rainbows may not occur if the droplets are too small. If the diameters are only equal to a few wavelengths of light, other optical effects prevent the development of rainbows.

The largest drops indicated in Table 9.2 represent the approximate upper limit to the size of raindrops. Deformation ultimately reaches the stage at which the drops are broken up into smaller droplets. There is therefore a limit to both the size and terminal velocity of raindrops.

The fall of hydrometeors towards the earth can only occur after clouds have developed in the sky. Clouds form when the cooling of water vapour reaches the stage where it changes into liquid droplets or solid ice crystals.

The cooling of moist air to form clouds involves vertical motion. In the next chapter we shall discuss some of the physical processes which lead to the upward motion of the air.

CHAPTER 10

THE VERTICAL STABILITY OF THE ATMOSPHERE

Cloud formation and precipitation are largely the result of vertical motions in the atmosphere. Sometimes these motions are visible to the eye as we watch the development of clouds in the sky. At other times, vertical motion may occur in the absence of clouds.

Vertical motion in the atmosphere is generally on a much smaller scale than horizontal motion. Nevertheless, when it is widespread or well developed, its effects can be very important.

Varied weather phenomena arise from vertical motion. Irregular local gusts and lulls may occur for a period of a few seconds. Strong updrafts and downdrafts may continue for several minutes during thunderstorms. On other occasions, widespread but slower movements may be sustained for days at a time, due to the effects of large scale weather systems.

These motions are related to the vertical stability of the atmosphere, and we shall now study how this arises.

10.1 Adiabatic processes in the atmosphere

An adiabatic process is one which takes place without exchange of heat with the environment. Thus a change in the volume or pressure of a small parcel of gas may occur without the flow of heat into or out of the parcel.

Many short-term atmospheric pressure changes are adiabatic or nearly adiabatic for three reasons. First, air is a poor conductor of heat. Second, mixing of a parcel of air with its surroundings usually takes place relatively slowly. Third, radiative processes produce only very small changes during short-term periods.

A parcel of unsaturated air may expand adiabatically when it rises and enters regions of lower pressure aloft. In doing so, it cools at the dry adiabatic lapse rate. This cooling due to adiabatic expansion is at the rate of about 10°C per kilometre, assuming that the air remains unsaturated.

The expression dry adiabatic lapse rate (DALR) is used, because the adiabatic temperature changes that occur in unsaturated air are very approximately equal to those in dry air. It is important to note, however, that the moist air must remain unsaturated if the dry adiabatic lapse rate is to be applied.

Sometimes, however, moist air may become saturated due to cooling. A different adiabatic lapse rate then applies, if the parcel of saturated air continues to rise and expand.

Adiabatic cooling of saturated air leads to the condensation of some of the water vapour and cloud begins to form. Latent heat is released and this partly counteracts the adiabatic cooling due to expansion.

If the air is saturated, the lapse rate is therefore smaller than the dry adiabatic lapse rate. Saturated air is said to cool at the saturated adiabatic lapse rate (SALR).

The value of the saturated adiabatic lapse rate depends upon the pressure and the temperature. The reason for the dependence on temperature is that air holds more moisture at higher temperatures. More latent heat is released and this reduces the rate of cooling.

It is therefore impossible to speak of a single saturated adiabatic lapse rate. Table 10.1 shows some of its values at different pressures and temperatures.

Table 10.1 Saturated adiabatic lapse rates

Temperature (°C)	Saturated adiabatic lapse rate (°C per km)	
	Pressure 1000 mb	500 mb
30	3.6	
20	4.5	
10	5.6	4.2
0	6.9	5.4
−10	8.1	6.8
−20	8.8	8.4

In some examples we shall use a value of 5°C per kilometre for the saturated adiabatic lapse rate. This is only for convenience, because the actual conditions must be considered in any particular situation.

10.2 The environment lapse rate

In Chapter 2 we dealt with the vertical divisions of the atmosphere based on temperature. These changes of temperature referred to mean conditions in the atmosphere.

Each day in any locality the vertical distribution of temperature with height varies from these average conditions. For this reason, meteorological observers launch balloons which carry radiosonde equipment high into the atmosphere. In this way, it is possible to determine the pressure, temperature and humidity at the various levels in the atmosphere.

A knowledge of the vertical distribution of temperature enables a meteorologist to determine the environment lapse rate at various levels. The environment lapse rate is the rate of decrease of temperature with height.

The environment lapse rate (ELR) is positive if the temperature decreases with height. By contrast, if the temperature increases with height through a layer of the atmosphere, the environment lapse is negative and a temperature inversion is said to occur.

The environment lapse rate may be greater than, equal to or less than the dry adiabatic lapse rate for a small parcel of unsaturated air moving vertically through the atmosphere. It may similarly differ from the saturated adiabatic lapse rate of a parcel of saturated air in vertical motion.

Differences between the environment lapse rate and the adiabatic lapse rates for parcels of air lead to important events in the atmosphere.

10.3 Stability

If the forces acting on an object are balanced, it may remain at rest. It is then said to be in a state of equilibrium.

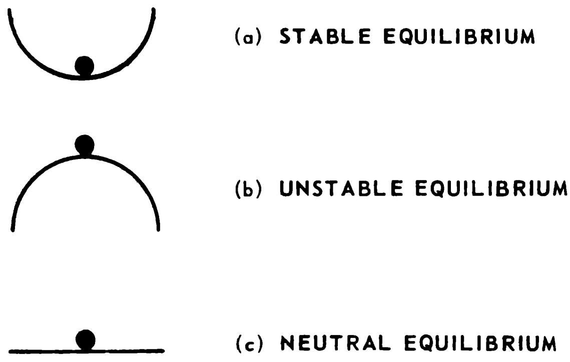
Objects in a state of equilibrium respond differently when they are subject to a small displacement from their equilibrium position. Some tend to return to the original position. In such a case, the object is said to be in stable equilibrium.

Others tend to move still further away from their original position. Such objects are considered to be in unstable equilibrium.

A third possibility may also occur. The object may tend neither to return to its original position nor to move further away; it remains in its new position. This is known as neutral equilibrium.

Figure 10.1 indicates the three different types of equilibrium of a spherical object placed on surfaces of different shapes.

Figure 10.1 States of equilibrium



10.4 **Parcel method stability**

One method of investigating whether vertical motion is likely to occur in the atmosphere is to consider the vertical displacement of a small parcel of air. It is first assumed that the parcel is in a state of equilibrium and the forces on it are balanced.

It is then given a small vertical displacement upwards. If it tends to return to its original position it must be in stable equilibrium. In these circumstances, the atmosphere is said to be stable. Note that in a stable atmosphere vertical motions are either absent or definitely restricted. For instance, the vertical development of clouds is limited in stable regions of the atmosphere.

In some situations an unstable atmosphere occurs. This will arise if a small parcel of air continues to rise still further after receiving a small upward displacement. Vertical motions are prevalent in an unstable atmosphere. If the air becomes saturated, clouds may extend to great heights.

A third possibility is a neutral atmosphere. There is then no tendency for the air to move either up or down from its new position.

In each of these three cases the direction of the vertical motion of the displaced parcel of air will depend on whether its temperature is higher or lower than that of its environment, i.e. the surrounding air.

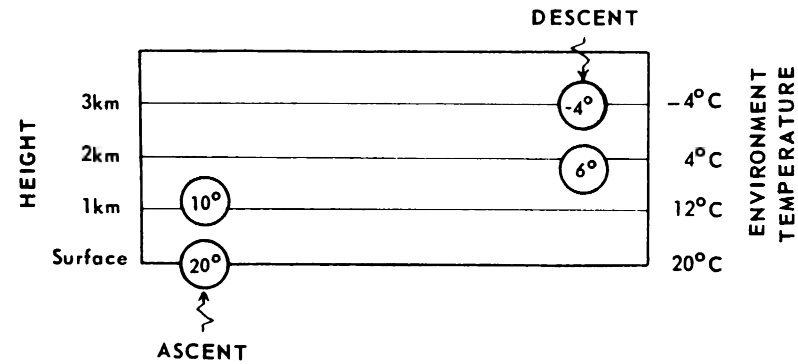
Whether the atmosphere is stable, unstable or neutral therefore depends on the temperature changes that occur with height. We shall now consider the situation for unsaturated and saturated air in turn.

10.5 The vertical motion of unsaturated air

Let us consider a portion of the atmosphere where the relative humidity is well below 100% and so the air is unsaturated. Suppose that the surface temperature is 20°C. Let us now consider three different environment lapse rates.

(a) Suppose the environment lapse rate (ELR) is 8°C per kilometre. Figure 10.2 shows the distribution of temperature to a height of 3 km above the earth's surface.

Figure 10.2 A stable atmosphere for unsaturated air



Consider now a parcel of air in a state of equilibrium near the earth's surface. The parcel temperature is 20°C, the same as the environment temperature.

Suppose it is now displaced upwards to a height of 1 km. If it expands and cools adiabatically, the cooling will be at the rate of about 10°C per kilometre, i.e. at the dry adiabatic lapse rate (DALR).

If you refer again to Figure 10.2, you will notice that at a height of 1 km the parcel temperature is 10°C, while the environment temperature is 12°C. The parcel is therefore 2°C colder than the surrounding air.

As it is denser than the environment at that level, the parcel of air will sink back to its original position. During the descent it will be compressed, and assuming that this occurs adiabatically, its temperature will rise 10°C to reach the original value of 20°C.

During the above events, the parcel of air was displaced upwards from its equilibrium position. It then returned to its original position and so it must be in stable equilibrium.

A stable atmosphere therefore occurs in the above situation. Notice that the environment lapse rate (ELR) is 8°C per kilometre. This differs from the rate of cooling of the parcel of air, which is 10°C per kilometre, i.e. the dry adiabatic lapse rate (DALR).

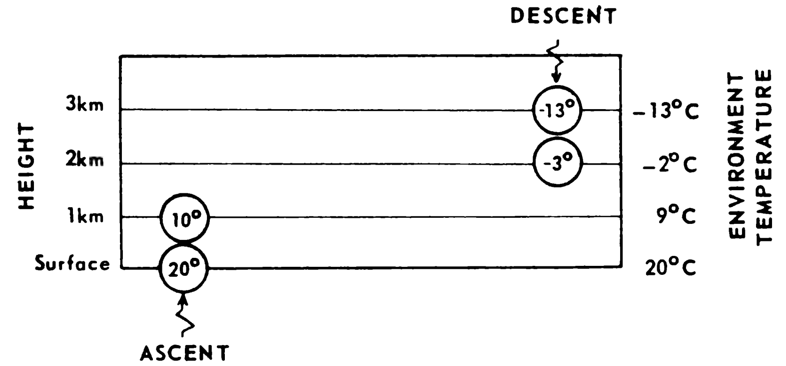
For a stable atmosphere,

$$\text{ELR} < \text{DALR} \qquad \ldots 10.1$$

if the parcel of air remains unsaturated.

(b) Suppose the environment lapse rate is 11°C per kilometre. Figure 10.3 shows the distribution of temperature to a height of 3 km above the earth's surface.

Figure 10.3 An unstable atmosphere for unsaturated air



In this case, the parcel of air again reaches the 1 km level at a temperature of 10°C. However, it is now 1°C warmer than its surroundings, because the environment temperature at 1 km is only 9°C.

As a result, it will be less dense than the surrounding air and will be forced to rise still further. This is therefore a case of unstable equilibrium. An unstable atmosphere therefore occurs in this situation.

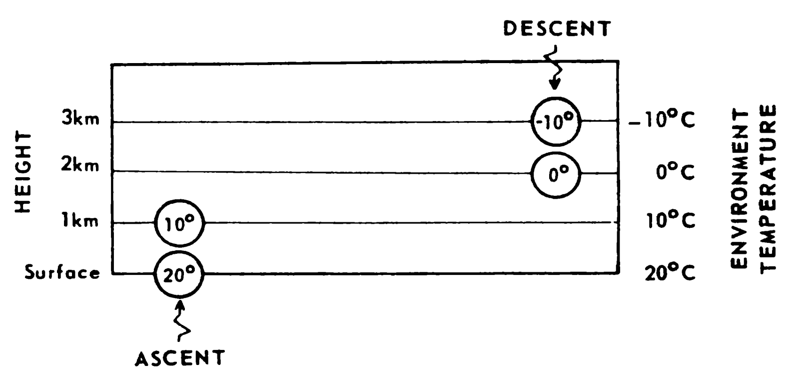
Notice that for an unstable atmosphere,

$$\text{ELR} > \text{DALR} \qquad \ldots 10.2$$

if the parcel of air remains unsaturated.

(c) Suppose the environment lapse is 10°C per kilometre. Figure 10.4 shows the situation in which the environment lapse rate is 10°C per kilometre, i.e. equal to the DALR. In this case, the displaced parcel of air at 1 km is at the same temperature as its environment. It therefore tends to remain in its new position. A neutral atmosphere therefore occurs under these conditions.

Figure 10.4 A neutral atmosphere for unsaturated air



For a neutral atmosphere,

$$\text{ELR} = \text{DALR} \qquad \ldots 10.3$$

if the parcel of air remains unsaturated.

In the above discussion we have only referred to a vertical displacement upwards, i.e. ascent. In the top corner of each diagram you will notice the temperature changes which occur during a displacement vertically downwards from 3 km to 2 km, i.e. descent.

During the descent the parcel of air will warm at the dry adiabatic lapse rate. You should check for yourself that the following events occur:

If ELR $<$ DALR the parcel returns to 3 km level (stable atmosphere)
ELR $>$ DALR the parcel continues to descend (unstable atmosphere)
ELR $=$ DALR the parcel remains at 2 km level (neutral atmosphere).

10.6 **The vertical motion of saturated air**

If a parcel of air is saturated, the release of latent heat reduces its rate of cooling when it expands adiabatically. We must therefore use the appropriate saturated adiabatic lapse rate (SALR) instead of the dry adiabatic lapse rate.

The value of the saturated adiabatic lapse rate varies with the temperature and pressure, but you may assume that it has a value of 5°C per kilometre in this exercise. Test for yourself what happens when the surface temperature is 20°C and the environment lapse rate (ELR) is 4°C per km.

First displace a parcel vertically upwards from the surface to a height of 1 km and study the effect. Later, investigate what happens to a parcel of saturated air if it is forced to descend from the 3 km to the 2 km level.

Repeat the exercise for ELR = 8°C per km and ELR = 5°C per km.

You should then be able to establish the following relationships (for a saturated parcel of air):

Stable atmosphere: ELR $<$ SALR
Unstable atmosphere: ELR $>$ SALR
Neutral atmosphere: ELR $=$ SALR.

10.7 **Conditional state**

If you consider the situation in which the environment lapse rate is 8°C per kilometre, you will note that

$$\text{ELR} < \text{DALR}$$
$$\text{but ELR} > \text{SALR}.$$

The atmosphere is therefore stable for unsaturated parcels of air. By contrast, it is unstable if they are saturated.

Such a condition is known as the conditional state. It occurs for an environment lapse rate such that

$$\text{SALR} < \text{ELR} < \text{DALR}.$$

The terms 'conditional stability' and 'conditional instability' are sometimes used to describe this situation, but have exactly the same meaning.

10.8 **Summary of vertical stability**

A knowledge of the stability of a parcel of air at rest is a useful indication of the possibility of vertical motion. Because the parcel is assumed to be initially at rest or stationary, this concept is often referred to as static stability.

In the preceding sections we have assumed that the parcel of air is given a small vertical displacement. It then moves into a region of different pressure and its volume changes. We then assume that no heat is added or subtracted from the parcel of air, i.e. it expands or contracts under adiabatic conditions.

In studying most atmospheric processes for a period of about a day it is a reasonable assumption that they are adiabatic for three reasons:

(a) air is a poor conductor
(b) mixing of the parcel with its surroundings is slow
(c) radiative processes only produce very small changes during short-term periods.

However, in some situations mixing of the air with its surroundings may be important. For instance, during the development of cumuliform clouds, such as cumulonimbus, the surrounding air may be drawn into the region of rising air. This is known as entrainment and mixing of the air will occur during this process.

Meteorologists use special methods of studying the stability of the atmosphere in regions where entrainment is important. Not only do heat transfers occur but also changes in the moisture content of the air may arise.

They also make allowances for non-adiabatic changes in temperature, if they wish to make predictions of vertical motion for periods longer than about twenty-four hours. Non-adiabatic processes include the gain or loss of heat by radiation.

In general, however, you will find that the parcel method of determining the static stability will provide a useful guide to the possibility of vertical motion. It will be most useful during the short-term periods, which follow the determination of the environment lapse rate from radiosonde observations.

We have noted that two properties largely determine the stability of the atmosphere during short-term processes:

(a) The change in temperature with height in the environment—the environment lapse rate (ELR)
(b) The change in temperature with height of a parcel of air as it ascends or descends.

In case (b) we use the dry adiabatic lapse rate (DALR) if the parcel of air remains unsaturated. If the air is saturated, it is necessary to use the appropriate saturated adiabatic lapse rate (SALR).

We may therefore summarise the possibilities as follows:

(a) If ELR $<$ SALR the atmosphere is always stable
(b) If SALR $<$ ELR $<$ DALR the atmosphere is stable for unsaturated air, but unstable for saturated air (i.e. conditional instability occurs)
(c) If ELR $>$ DALR the atmosphere is always unstable.

In cases where the ELR equals the SALR or DALR, the possibility of neutral stability should be investigated.

10.9 **The lifting condensation level**

Let us consider a small parcel of unsaturated air which is forced to rise in the atmosphere. As it rises, it cools at the dry adiabatic lapse rate (DALR). If the motion continues, the parcel will cool to the stage where it becomes saturated.

Any further cooling will cause condensation (i.e. cloud formation) to commence. This level is called the lifting condensation level.

If we study the motion of the parcel of air above the lifting condensation level, we must consider it cooling at a slower rate. This is due to the fact that the release of latent heat partly counteracts the adiabatic cooling. We then compare the saturation adiabatic lapse rate (SALR) with the environment lapse rate (ELR) determined from a radiosonde observation.

10.10 **Atmospheric turbulence**

If you examine an anemograph trace you will notice that the surface wind usually undergoes rapid and irregular fluctuations in both speed and direction. These fluctuations indicate that the air flow is turbulent with many eddies occurring in the region near the earth's surface.

It is difficult to determine the exact structure of these eddies but they are very irregular. In particular, their axes of rotation may be inclined in any direction.

The degree of turbulence has been found to depend on a number of factors, such as the speed of the wind, the roughness of the surface, the environment lapse rate, etc.

Meteorologists often distinguish two types of turbulence:

(a) Thermal turbulence
(b) Mechanical turbulence.

Thermal turbulence results from the convection currents set up by surface heating. This heating may result from insolation on the land. Sometimes it also occurs when a relatively cool mass of air passes over a warmer land or sea surface.

As the temperature rises near the earth's surface the environment lapse rate increases. Eventually, the air becomes unstable and convection currents become part of the turbulent structure.

In convective clouds, especially in thunderstorms, latent heat is released at cloud heights and this energy sets up updrafts and downdrafts, representing large eddy-type motions. These in turn give rise to numerous smaller turbulent eddies of various sizes.

Thermal turbulence or convection does not always lead to the development of clouds. In hot, arid regions the humidity of the atmosphere may be too low for condensation to occur. Turbulence may nevertheless be very great in these regions. Aircraft are affected by the turbulence and this is known as 'bumpiness'. In some localities, the effects of turbulence are noticeable if dust storms develop.

The other important type of turbulence is known as mechanical turbulence. This is also known as frictional turbulence and is widespread due to the roughness of the earth's surface. Turbulence is accentuated by the flow of air over buildings, trees, hills, etc. and is also generated by wind shear.

Fluid actually in contact with a stationary boundary is itself at rest. However, throughout the friction layer the mean speed increases with distance from the boundary until the unretarded free stream is reached. Smooth or laminar flow occurs if the speed of the free stream is sufficiently low for a given fluid and type of boundary.

However, when the free speed exceeds a certain limit the flow breaks down into turbulent motion. Eddies then form near the boundary and drift away into the stream. The friction layer becomes much deeper in turbulent than in laminar flow.

The eddies produced by mechanical turbulence may rotate about axes in any direction. They tend to develop more easily if the wind speed is high or if the environment lapse rate is great.

Mechanical turbulence tends to be lower over the open sea or relatively smooth ground. Light winds or calm conditions and a stable atmosphere are also conducive to reduced mechanical turbulence.

A severe form of turbulence may occur in clear air at high levels in the atmosphere. This is known as clear air turbulence. It often presents a hazard to aircraft flying in these regions.

The effects of the various types of atmospheric turbulence will be discussed in more detail at a later stage.

10.11 **Temperature inversions**

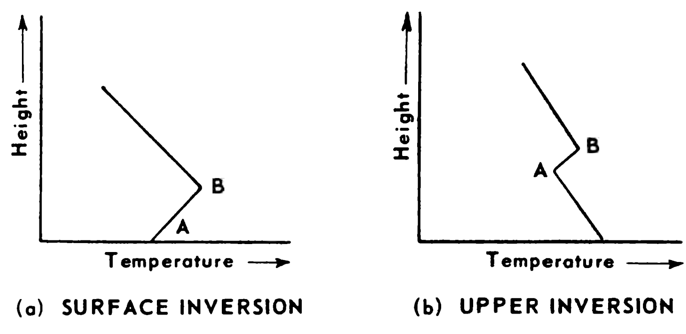
The environment temperature usually decreases with height in the troposphere. However, the temperature sometimes increases with height in certain layers. This is called a temperature inversion or simply an inversion.

An inversion may sometimes occur from the ground level upwards. This is known as a surface inversion.

On other occasions an inversion occurs in a layer situated above the earth's surface. This is called an upper inversion.

Figure 10.5 illustrates these two effects. The point A is the base of the inversion and the point B represents the top. In the case of the surface inversion the base is at ground level.

Figure 10.5 Surface and upper inversions



Temperature inversions may be produced in a variety of ways. The following four processes are important:

(a) Radiation
(b) Turbulence
(c) Subsidence
(d) Frontal development.

10.12 **Radiation inversions**

A radiation inversion may produce a surface inversion. At night the surface of the earth cools by radiation. If this cooling continues long enough the air near the earth's surface becomes cooler than the atmosphere above. A surface inversion then develops near the ground (see Figure 10.5(a)).

With calm or very light winds the cooling of the air extends upwards through a comparatively small height. This is called a shallow inversion.

The temperature at the surface may, however, be quite low and the shallow surface inversion may be quite marked. It is then described as a strong inversion. On cloudless nights with little wind, radiation inversions occur. Early morning fogs (radiation fogs) may then develop if the air contains sufficient moisture.

In some situations, frosts may occur. In these cases, the air is less humid, and radiation may occur more rapidly. Hence, lower surface temperatures develop, especially in inland regions after a long cloudless winter night.

Radiation also occurs from the tops of clouds at night. In this way, a radiation inversion may develop in the atmosphere well above the earth's surface.

10.13 **Turbulence inversions**

Turbulence often contributes to the development of inversions. If continued long enough, a thorough mixing of the atmosphere takes place in the layers where the turbulence exists.

Thus, mechanical turbulence may cause the cold air at the bottom of a surface inversion to be carried aloft. The cooling produced by radiation cooling may therefore be spread through a thicker layer of air. The top of the inversion therefore occurs at a greater height.

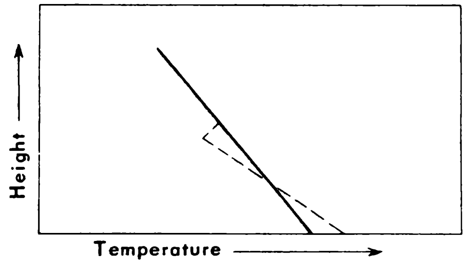
On the other hand, if the wind is stronger the turbulence may be even greater. The mixing then causes the cooler air to be spread through a much thicker layer. As a result, there is little lowering of the temperature, and so an inversion does not occur. This indicates that the wind strength and the consequent turbulence must lie between certain limits if a deep surface inversion is to occur.

Turbulence may sometimes produce an upper inversion. In the turbulence layers air is brought downward and heated by adiabatic compression. At the same time, air from lower levels is raised and cools as a result of adiabatic expansion.

After some time, all the air in the turbulence layer will have undergone adiabatic expansion and compression in this mixing process. An adiabatic lapse rate will develop within the layer; air at the bottom will be warmer than formerly, while that at the top will be colder.

Above the turbulence layer the temperature will not be affected by adiabatic cooling. As a result a turbulence inversion will be produced. Figure 10.6 illustrates these changes, the broken line indicating the new temperature distribution.

Figure 10.6 Development of a turbulence inversion

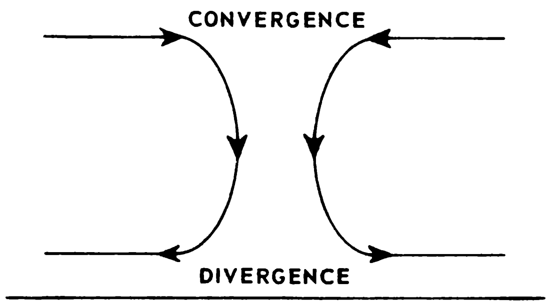


10.14 Subsidence inversions

In some regions of the atmosphere whole layers of air many hundreds of metres thick may sink or subside. This process may occur over a wide area and is known as subsidence.

This effect is associated with horizontal mass convergence and divergence, which were discussed in Section 7.14. Often convergence takes place in the upper troposphere at the same time as divergence occurs near the earth's surface. Figure 10.7 indicates how subsidence may result from these processes.

Figure 10.7 Subsidence associated with convergence and divergence



As some of the air flows outwards near the earth's surface, it is replaced by other air sinking downwards from aloft. The downward vertical velocity is greatest at about the middle of the troposphere. At higher levels beneath the tropopause, the air flows inwards before it makes its way downwards to lower levels.

As the air spreads outwards near the earth's surface the thickness of a layer of subsiding air generally decreases. In this case the top of the layer sinks more than the base.

Now subsiding air warms because it undergoes adiabatic compression as it reaches the higher pressure regions near the earth's surface. If the top of the layer subsides more than the base, it will be warmed more than the base. If the top of the subsiding layer reaches a higher temperature than its base, a subsidence inversion is formed.

Subsidence is associated with high pressure areas (anticyclones). Convergence aloft may cause pressures to rise near the earth's surface. Low-level divergence may then occur initially as the pressure gradient force drives the air outwards. Gradually, however, the Coriolis force increases with the wind speed and tends to balance the pressure gradient force.

The air does not, however, flow exactly around the isobars. The friction force produces outward cross-isobar flow, and this also contributes to the low level divergence.

10.15 Frontal inversions

In Chapter 16 we shall discuss the frontal zones which separate air masses possessing different densities and temperatures. If warm air is forced to rise over cool air in the vicinity of a frontal zone, the conditions necessary for the formation of an inversion are present. This is known as a frontal inversion.

Fronts and frontal inversions will be discussed in more detail at a later stage. However, it should be mentioned that in addition to a rise in temperature, there is sometimes an increase in the water vapour content at a frontal inversion. In this way, it may differ from other types of inversions, in which a rapid decrease in moisture content usually accompanies the temperature rise.

10.16 The effects of inversions

The stability of the air in an inversion is very great. Vertical motion is therefore resisted and tends to die out rapidly.

The existence of an inversion is often indicated by the tops of clouds that are prevented from extending upwards. Their tops therefore spread out beneath the base of the inversion layer.

Similarly, haze, caused by smoke or dust, is often confined below a temperature inversion.

Optical effects are also associated with inversions. The refraction or bending of light rays depends on the temperature and water content of the air. In the vicinity of temperature inversions, refraction may be abnormal. Some mirages arise when light rays are bent as they travel downwards from a warm layer of air to a colder one below.

The transmission of radio waves of wavelengths less than 10 m are similarly affected by changes in the temperature and water vapour content of different layers of air. Temperature inversions and the associated differences in the densities of various layers of air are particularly important due to their effects on the paths of the shorter radio waves used in microwave transmissions and in radar.

10.17 Superadiabatic lapse rates

Lapse rates greater than the dry adiabatic are called superadiabatic. The atmosphere is then unstable and vertical currents usually develop. These mix the air and redistribute the heat, until the environment lapse rate becomes equal to the dry adiabatic lapse rate.

Superadiabatic lapse rates are however sometimes found within a layer that extends a few metres upwards from the ground. This situation may occur when the ground is exposed to strong sunshine, and light winds or calm conditions prevail.

10.18 Convection condensation level

The environment lapse rate near the earth's surface undergoes changes during the day-time and night-time. We have already investigated changes leading to the formation of temperature inversions.

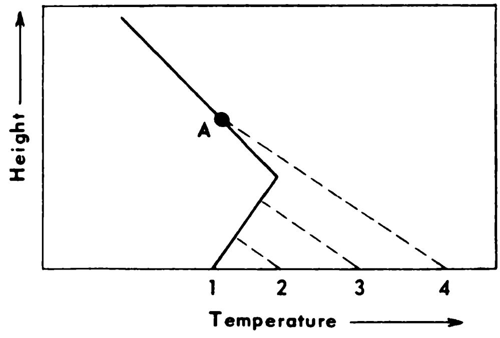
Other effects arise due to solar radiation heating the earth's surface. As the day advances, this heat is transferred from the ground to the lower layers of the atmosphere. The depth of air heated is indicated by the thickness of the layer, through which the environment lapse rate becomes equal to the dry adiabatic lapse rate.

Figure 10.8 indicates how the temperature distribution changes as the dry adiabatic lapse rate becomes established through deeper and deeper layers.

As the temperature rises during the day-time, a stage may eventually be reached when the dry adiabatic lapse rate extends up to the level at which condensation takes place. This is indicated by the point A in Figure 10.8.

This is known as the convection condensation level. The temperature at the surface is then called the convection temperature.

Figure 10.8 Convection condensation level



When the surface temperature reaches the convection temperature, and the dry adiabatic lapse rate is extended up to the condensation level, cumulus cloud will develop. Its base will occur at about this level.

We have studied the static stability of the atmosphere and its relation to vertical motion. This in turn leads to the development of clouds if the motion continues until condensation of the water vapour occurs. In the next chapter we shall investigate the physical processes of cloud formation in more detail. In addition, we study the various factors which lead to their dispersal.

CHAPTER 11

CLOUD FORMATION AND DISPERSAL

Although water is present in some degree almost everywhere in the troposphere, it is usually unseen in the form of vapour. Whenever clouds appear they provide visual evidence of the presence of water in the atmosphere. In many cases they indicate future trends in the weather.

A mixture of dry air and water vapour is known as moist air. Most clouds are formed by the cooling of moist air. Therefore those atmospheric processes that produce cooling of the air may also result in the formation of clouds.

In this chapter we shall first study the various ways in which clouds are formed in the atmosphere. Later, we shall investigate what happens when these processes cease and the clouds begin to disperse.

11.1 Condensation, freezing and deposition

In Chapter 6 we discussed the processes by which water vapour may change to liquid water droplets or to solid ice crystals.

When moist air is cooled below its dew-point, water droplets condense on the condensation nuclei present in the air. In some cases these nuclei have a special affinity for water and are said to be hygroscopic. Salt particles from sea spray are of this type and condensation may occur before the relative humidity reaches 100%.

When water droplets in the atmosphere are cooled below 0° C they do not necessarily freeze. Under such conditions they are said to be supercooled. Cloud droplets commonly occur in the supercooled state at temperatures as low as -20°C. On occasion, supercooled cloud droplets may even occur at temperatures down to -35°C.

Laboratory experiments have shown that small droplets may be cooled to temperatures of about —40°C before freezing occurs. However, it is found that water in contact with foreign solid materials or small suspended particles will freeze at considerably higher temperatures.

In the atmosphere, certain types of suspended particles may act as nuclei for the freezing process. A particle which initiates the growth of an ice crystal by the freezing of supercooled water about itself is called a freezing nucleus.

Water vapour may also change directly to a solid without passing through the liquid state. This process is known as deposition. It is also sometimes referred to as sublimation, but this name really refers to the reverse process, in which a direct change from ice to water vapour occurs.

Any particle upon which an ice crystal grows by the process of deposition is called a sublimation nucleus. Many experiments have been conducted to determine whether certain particles act as freezing nuclei or sublimation nuclei. It has never been demonstrated that sublimation nuclei, as distinct from freezing nuclei, exist in the atmosphere.

In meteorology it is therefore usual to use the term freezing nuclei when referring to ice-forming nuclei. A thin film of water first forms on the surface of the nucleus and then freezes. The film is so thin, however, that droplets are difficult to detect. As a result, the ice crystal forms as if from a vapour.

Ice-forming nuclei are also referred to as ice nuclei. The fact that water droplets occur so frequently in clouds at temperatures below 0°C indicates that ice nuclei are rarer than condensation nuclei. Nuclei which cause the formation of ice crystals in clouds at temperatures above −40°C are not the same particles as those responsible for the condensation of water vapour to cloud droplets.

The majority of freezing nuclei in the atmosphere probably originate at the earth's surface as wind-blown soil particles of particular types. Certain clay minerals seem to be the most important, and it is likely that turbulent mixing could distribute them fairly uniformly up to high altitudes.

11.2 General causes of cloud formation

Most clouds are formed when moist air is subjected to upward motion. As it rises it cools as the result of the expansion which takes place at the lower pressures in the upper atmosphere. Some of the water vapour then condenses to form a cloud.

In Chapter 8 we discussed the characteristics of cloud genera. The shapes and forms of these clouds are really expressions of the way in which the air has risen to fashion them. The various types of vertical motion that lead to the formation of clouds are as follows:

(a) Mechanical turbulence (or frictional turbulence)
(b) Convection (or thermal turbulence)
(c) Orographic ascent
(d) Slow widespread ascent.

11.3 Mechanical turbulence

In Section 10.10 we discussed mechanical or frictional turbulence. Air flow over the earth's surface is generally deformed by frictional forces into a series of eddies. This turbulent motion is accentuated by buildings, trees, hills, etc.

We mentioned in Section 10.13 that air in the friction layer is thoroughly mixed by mechanical turbulence. If the layer is initially stable, the upper part will be cooled and the lower portion warmed. As a result, a dry adiabatic lapse rate may become established in the layer if the air remains unsaturated.

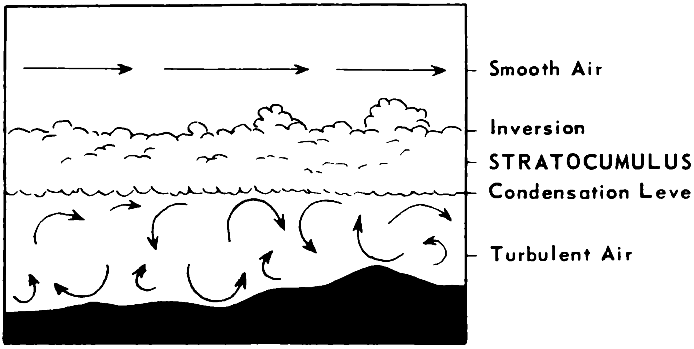
Similarly, the turbulence mixes the water vapour in the turbulent layer. The water vapour content in the layer therefore tends to become evened out. As a result, the air may become saturated at some distance below the top of the friction layer. Condensation may then occur at a height above the ground known as the mixing condensation level (MCL). This will represent the base of the cloud.

If cloud forms due to turbulence, the dry adiabatic lapse rate will only extend up to its base (MCL). The saturated adiabatic lapse will then extend to the top of the turbulent layer. The cloud extends upwards to the base of the turbulence inversion at the upper limit of the friction layer.

The cloud formed by turbulence is initially stratus, a sheet cloud without definite form. It may either continue as such, or else the upper and lower surface may develop a wave-like appearance.

When these undulations occur, the thickness of the cloud may then vary, and sometimes breaks may be seen. These arise as a result of cloud being formed in the up currents, and evaporated in the down currents. This cloud is classed as strato-cumulus and its development is shown in Figure 11.1.

Figure 11.1 Formation of turbulence cloud



Turbulence clouds may also develop below rain-bearing clouds, such as nimbo-stratus, altostratus and cumulonimbus. They are the very low ragged clouds of bad weather, known as stratus fractus or cumulus fractus.

The word 'fractus' means 'broken'. Sometimes the prefix 'fracto' is used with the same meaning. Thus, the low broken clouds are referred to as fracto-stratus and fracto-cumulus.

These clouds derive their moisture from the evaporation of raindrops and surface rain water. Turbulence near the earth's surface then produces ragged low clouds in the air beneath the main cloud formation.

Sometimes high stratocumulus or altocumulus are observed when the wind changes with height through a humid layer, high above the friction layer. Turbulent motion may then occur at these heights, but usually some other factor is responsible for the high water vapour content. The moisture has not been transferred to that level by direct mixing due to turbulence near the earth's surface.

11.4 **Convection**

When air is heated near the surface, convection currents develop. This process is known as convection or thermal turbulence. It combines with mechanical or frictional turbulence to mix the air in the lower layers of the atmosphere.

In general, the environment lapse rate tends to the dry adiabatic lapse rate, while the air remains unsaturated. In Section 10.18 we mentioned that this lapse rate may be established up to the convection condensation level.

Cloud forms at this level, but the extent of its upward development depends on a number of factors. One is the environment lapse rate in the air above the cloud base.

If the environment lapse rate is greater than the saturation adiabatic lapse rate, the atmosphere is unstable for saturated parcels of air. The saturated air is then able to rise freely. It continues to rise until it reaches a level where it is no longer warmer than its surroundings.

A cumuliform cloud develops in this way. The distance from the base to the top may vary from one or two kilometres to ten or more kilometres. Isolated convection clouds of limited extent are called fair-weather cumulus. Their vertical development is insufficient for precipitation to occur.

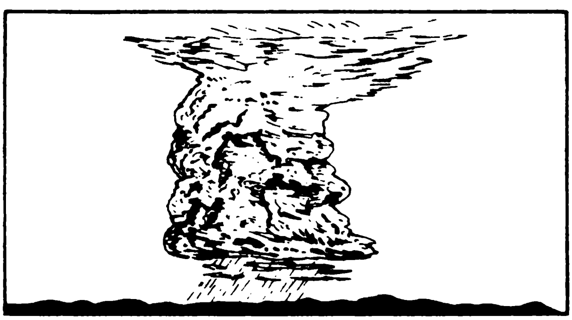
Sometimes the vertical currents are terminated by a marked inversion above the convection condensation level. The top of the cloud then spreads out beneath the inversion and the cloud develops into stratocumulus.

In some situations the environment lapse rate above the condensation level exceeds the saturation adiabatic lapse rate through a deep layer. If sufficient moisture is available the cloud may then extend to great heights.

The tops of the clouds may sometimes reach levels where ice crystals form. A convection cloud of great vertical extent and whose top is composed of ice crystals is known as cumulonimbus—the thunderstorm cloud. The veil of ice crystals that surrounds the upper parts of the cloud gives it a smooth fibrous appearance, which distinguishes it from a cumulus cloud.

Figure 11.2 illustrates the appearance of cumulonimbus cloud with a fibrous or striated upper part present.

Figure 11.2 Cumulonimbus cloud with fibrous top



Slight or heavy precipitation may develop according to the degree of instability and to the height and temperature reached. The cloud may sometimes have a base less than a kilometre and a vertical extent of 10 km or more. During showers, ragged turbulence cloud may develop below the main base, occasionally reaching almost to the earth's surface.

On reaching a stable layer or inversion the top of the cloud may spread out horizontally. The well-known anvil shape of the top of a cumulonimbus cloud develops in this way.

When instability conditions are well-developed, enormous amounts of energy become available from the release of latent heat. Vertical velocities in up-currents may exceed 10 m/s, and temporarily prevent even the largest raindrops from falling downwards. Later, they may be released if the vertical currents are interrupted. A violent rainstorm or cloudburst may then occur.

Cumulonimbus clouds often reach a greater height in the tropics than elsewhere. The greater height is permitted by the high tropopause and cloud tops extending to a height of 16 km or more are not uncommon.

In the tropics the temperature at the condensation level is higher, and consequently the water vapour content of the air is higher. Condensation therefore releases greater quantities of latent heat, and thunderstorms of great violence occur.

Phenomena associated with convection are of great importance in the tropics. Further aspects will be discussed in later chapters.

11.5 **Orographic ascent**

Orography concerns the study of mountains. In this section we shall consider what happens when moist air ascends over mountain ranges or a barrier of hills.

When the air reaches the barrier, it is forced to rise both near the surface and at upper levels. A deep layer of the atmosphere may be affected and the vertical distribution of temperature will be altered. The air that has been forced to ascend cools adiabatically, and clouds may form.

The type of cloud formed by orographic ascent depends on a number of factors. One of these is the stability of the air in which it is formed.

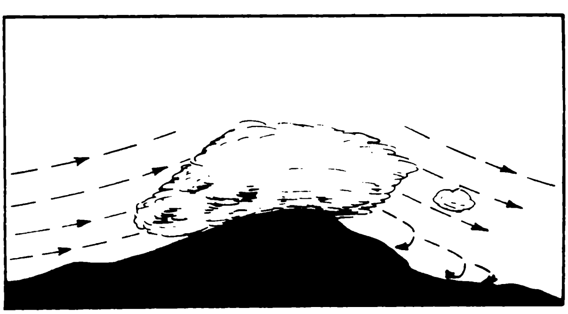
In moist stable air stratus is frequently formed. Cumulus cloud is characteristic of air which is slightly unstable. If instability is established through a great depth of the atmosphere, cumulonimbus cloud may develop.

Not all winds blowing on to a hill or mountain produce clouds. In some cases, the moisture content of the air is insufficient for cloud formation.

When stratus is formed it consists of a flat base and generally no great vertical thickness. It forms a sheet covering the higher ground, but breaks occur over lower-lying areas. The descent on the lee side causes the air to warm and so the cloud dissolves rapidly.

In general, orographic cloud forms continuously on the windward side of the hill or mountain, but clears on the lee side. The cloud as a whole appears to remain stationary, but the air itself actually continues on its way to the other side of the obstruction. Figure 11.3 illustrates the formation of orographic clouds.

Figure 11.3 The formation of orographic cloud

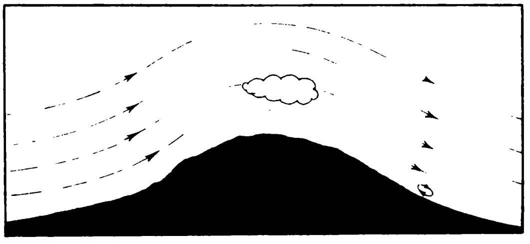


Sometimes cloud forms high above the hill or mountain. When there is a layer of almost saturated air aloft, orographic lifting of the air may cause condensation to occur above the obstruction. A persistent cloud cap may form.

Viewed from below, it is thin at the ends with a thicker and broader centre. Its shape is therefore similar to that of a lens. It is called a lenticular cloud. Figure 11.4 illustrates its formation.

Lenticular clouds appear to be stationary, just like the orographic clouds that form on the tops of the mountains. Actually, the water molecules and the other gases of the atmosphere stream through the cloud. At one end the water molecules condense to form cloud; at the other end they evaporate, changing back to the vapour state as they leave the cloud.

Figure 11.4 The formation of lenticular cloud



Sometimes a series of stationary waves may be formed in the lee of a range of hills. Under certain conditions of stability, the up-motion induced by the barrier increases with height. Downstream the air descends, sometimes through a greater distance than it initially ascended.

Later again, still further downstream, the air ascends once more. This process is repeated many times, gradually decreasing with distance from the barrier. The particles of air therefore follow an undulatory path. Standing waves are said to occur.

Conditions are most suitable for waves when a very stable layer lies between an unstable layer at the surface and another above. The waves often have their greatest amplitude within the stable layer.

If the moisture content and the amplitude of the undulations are sufficiently large, condensation takes place in the ascending air near the crests of the undulations; evaporation occurs in the descending air downwind of the crests. A series of lenticular clouds may then develop.

Lenticular clouds therefore indicate the presence of standing waves in the lee of mountain barriers. It is important to realise that the undulatory flow may still be present, even if the clouds are absent due to the low moisture content of the air. These standing waves frequently affect the performance of aircraft.

11.6 **Slow widespread ascent**

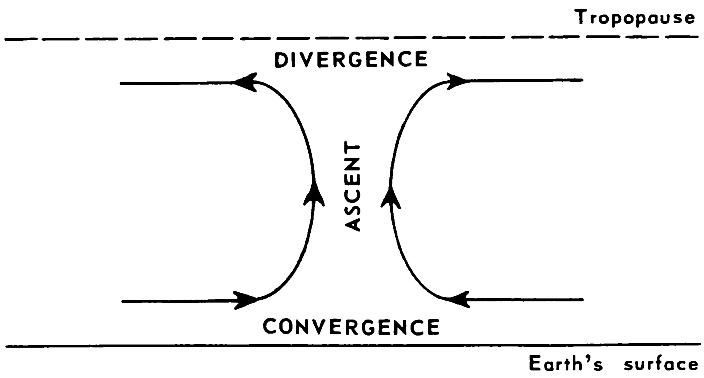
So far, we have discussed cloud formation arising from vertical motions associated with disturbances of relatively small horizontal dimensions. These vertical motions have been produced by frictional turbulence, local convection currents and orographic barriers, usually covering areas of a few kilometres in extent.

Vertical motion is also produced by large wind systems, such as the depressions (lows) and anticyclones (highs) located on MSL synoptic pressure charts. In Section 10.14 we studied the slow widespread descent of air that occurs in anticyclones. This is known as subsidence and may occur with high level convergence and low level divergence. The friction force is also responsible for some outward cross-isobar flow of the air near the earth's surface.

The reverse effect may be associated with depressions (lows). High level divergence and low level convergence can lead to ascent of the air as shown in Figure 11.5. The friction force may also produce a certain amount of low level convergence in the friction layer. This is associated with cross-isobar flow towards the centre of the depression.

The upward motion in a depression is distributed over a very extensive area and so the vertical velocities are relatively small. Nevertheless, the ascent may persist for many days causing large masses of air to ascend through many kilometres.

Figure 11.5 Ascent of air in a depression



Slow widespread ascent may have a marked effect on the environment lapse rate. The lapse rate increases and is said to steepen.

The air frequently becomes unstable, leading to increased vertical motion. Condensation and widespread cloud formation may occur, if the moisture content of the air is sufficiently high.

The cloud masses may sometimes be many kilometres in thickness. On the other hand, variations in the relative humidity may result in the formation of separate cloud layers.

Frequently, widespread ascent is first triggered by divergence in the upper troposphere. The resulting outflow of mass at high altitudes leads to a fall in pressure at lower levels in the vicinity of the earth's surface, and a depression is formed. Convergence then occurs near sea level and slow widespread ascent takes place through great depths of the troposphere. Widespread cloud development then follows, if the air is sufficiently moist.

Depressions and widespread ascent frequently occur in the vicinity of a frontal zone. This is a region which separates two large air masses differing in such properties as density and temperature. Depressions associated with frontal zones are called frontal depressions.

In a frontal zone there may frequently be a marked temperature change in a horizontal distance of a few hundred metres. In each air mass the temperature may be nearly uniform, but differ from the temperature of the air on the other side of the frontal zone.

Sometimes a frontal zone is referred to simply as a front. Remember however that in the atmosphere a sharp boundary does not occur between the air masses. A transition zone exists in which there is a change from the temperature of one air mass to that of the other.

11.7 Clouds associated with widespread ascent at frontal zones

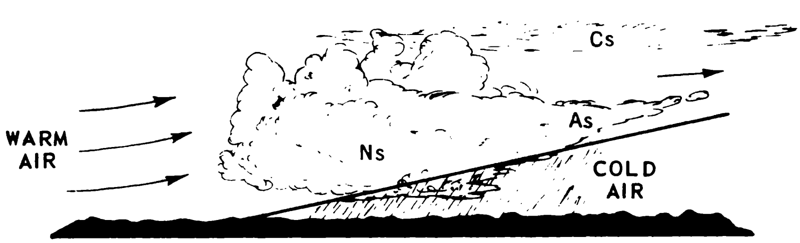
In Chapter 16 we shall discuss the polar front, which is one of the features of the general circulation. Cyclones develop along the polar front, which separates warm tropical air from colder air of higher latitudes.

Widespread ascent of air through deep layers of the atmosphere is frequently associated with polar front depressions. Many different relationships occur between the warm and cold air masses.

As the polar front cyclone develops, it frequently becomes possible to distinguish two main types of front—a cold front and a warm front. In each case, the frontal surface slopes upwards over the cooler air mass with the warmer air located above.

When the motion of the transition zone between the two air masses is such that warmer air replaces colder air it is called a warm front. The slope of a warm front is usually a gentle one and the warm air flows slowly upwards over the cold air mass. As it does so, stratiform clouds may develop in the warm air if the moisture content of the air is sufficiently high. Nimbostratus, altostratus, cirrostratus and cirrus clouds may develop at different étages in the atmosphere. Figure 11.6 shows this effect in an idealised situation.

Figure 11.6 Idealised cloud formation associated with warm fronts



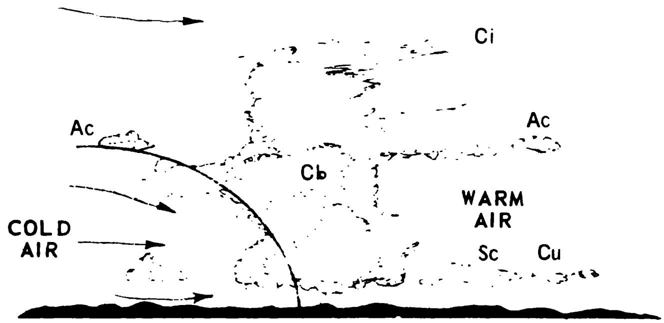
When the transition zone between the air masses moves in such a way that cold air replaces the warm air a cold front occurs. The cloud formation associated with a cold front varies with the stability and moisture content of the warm air, together with the slope of the front. On the average, the slope of a cold front is greater than that of a warm front.

If the slope of the cold front is weak, the cloud formation may be similar to that of a warm front, but with the various types of cloud arriving at a given locality in the reverse order. That is, the low clouds appear first, followed by higher stratiform clouds as the front moves away. The actual cloud formation in the warm air depends on the stability and moisture content of the ascending air.

Sometimes the slope of a cold front may be relatively steep. In this case, it may produce violent effects, particularly if the lifted warm air is already moist and unstable. It may then be characterised by the development of large cumulus and cumulonimbus clouds in the warm air. Heavy showers, gusty turbulent winds and sometimes thunderstorms may occur.

In general, the ascent of the warm air takes place within a narrower zone due to the steeper slope of the front. As a result, the cloud and weather phenomena of the steep-sloped cold front are usually confined to a narrower region in the vicinity of the frontal zone. Figure 11.7 shows the cloud development of an idealised cold front with a steep slope.

11.7 Idealised cold front with a steep slope



Many other types of cloud formation may occur, depending on the characteristics of both the warm and cold air, and on the stage of development of the polar front depression. These will be discussed further in later chapters.

11.8 Cloud dispersal

Cloud development will naturally begin to weaken once any of the processes leading to its formation cease to operate. Other factors may also lead to the disappearance of the water droplets or ice crystals from clouds. These include warming of the air, precipitation and mixing with drier surrounding air.

A cloud may be warmed by absorbing solar or terrestrial radiation. However, both these influences are relatively small, compared with adiabatic warming.

This may occur if the air in which the cloud exists is subjected to subsidence. As the temperature of the air rises, its relative humidity becomes lower and the air may no longer be saturated. Cloud particles then evaporate or sublime to form invisible water vapour.

Insolation often leads to the dissipation of cloud formed by turbulence. If sufficient solar radiation penetrates into the ground and so raises the air temperature near the surface, the mixing condensation level rises. The base of the stratus or stratocumulus rises, and the thickness of cloud beneath the turbulence inversion decreases. Eventually, the cloud may disappear completely.

Fair weather cumulus formed over the land by convection due to insolation is a day-time phenomenon. It usually appears in the morning, reaches a maximum development during the afternoon and clears rapidly when the ground cools towards evening.

The air surrounding a cloud is often unsaturated. Mixing of the cloud with this air may therefore result in the production of relative humidities well below 100%. Evaporation or sublimation then occurs, and the cloud appears to be eroded away.

The process of mixing with surrounding air markedly affects the base of fair-weather cumulus. The erosion of these clouds is most apparent at the base, and soon after their formation the cloud base becomes higher. In inland regions the maximum development of fair-weather cumulus is in the early afternoon. At this stage the base has not risen to any great extent, while the tops are near their maximum height.

Clouds are necessary for the development of precipitation. In Chapter 9 we studied the various forms in which precipitation might occur. Let us now investigate some of the processes which lead to precipitation.

CHAPTER 12

PRECIPITATION PROCESSES

It is necessary for water vapour in the atmosphere to change into water droplets or ice crystals before clouds can appear. These cloud particles must, however, increase still further in size before precipitation can take place.

In this chapter we shall consider the processes by which cloud particles grow to the stage where precipitation can occur. We shall also be concerned with the circumstances which lead to the different types of precipitation from clouds.

12.1 The sizes of cloud droplets

Meteorologists have expended a great deal of effort to obtain measurements of the sizes of water droplets in clouds. Most investigations have been made from aeroplanes, but studies have also been conducted in mountain areas.

A wide range of droplet sizes have been found in investigations carried out in different parts of the world. These droplets form on condensation nuclei of various sizes, types and concentrations.

Particles suspended in the atmosphere are often classified according to their size as:

(a) Aitken nuclei ($<0.1\ \mu$)
(b) Large nuclei (0.1 to 1.0 μ)
(c) Giant nuclei ($>1.0\ \mu$).

The figures in brackets indicate the approximate ranges of their radii.

The majority of Aitken nuclei are very small and require considerable supersaturation to become active in condensation. In the atmosphere the large and giant nuclei therefore lay first claim to the available water. The large nuclei are much more numerous than the giant nuclei and play an important role in cloud formation.

Cloud droplets compete for the available water vapour. If the concentration of condensation nuclei is high, the cloud droplets may be more numerous, but their average size could be smaller.

In general, greater concentrations of condensation nuclei occur above continents than over the oceans. On the average, droplets in continental clouds are smaller with typical radii ranging from 2 to 10 μ. On the other hand, maritime cloud droplets typically range from 3 to 22 μ in radius.

However, in almost all conditions giant salt nuclei occur. These may produce drops in the 20 to 30 μ range or larger. Their concentrations are often only about one per litre, but they occur in both maritime and continental clouds.

Cloud droplets may eventually reach such a size that they fall out of the clouds and the ascending currents which sustain them. An arbitrary division is sometimes made between cloud droplets and raindrops at a radius of 100 μ. This is usually a matter of convenience, but has a physical basis. The terminal velocity of drops of 100 μ radius is nearly 1 m/s, and so they are able to fall against the types of updrafts which normally occur in clouds.

In the next two sections we shall discuss processes by which cloud droplets grow to precipitation size.

12.2 The initial growth of cloud droplets

Many factors affect the growth of cloud droplets. These include the humidity of the air around the droplets, the effects of surface tension and the nature of the nuclei. The rapidity with which the released latent heat can be transferred to the surrounding air is also important.

At first, condensation of the water vapour on the nucleus proceeds very rapidly. The droplet may grow from the size of the nucleus to visible size in a fraction of a second. However, the process then slows down. Condensation alone is unlikely to produce droplets of average size greater than 30 μ.

If we are to explain the presence of larger drops within clouds, we must consider the interactions between individual droplets. An important method of growth occurs by the collision and fusing together (i.e. coalescence) of droplets falling relative to one another. We shall now consider this process in the next section.

12.3 The coalescence mechanism

An important process by which cloud droplets grow to raindrop size is the direct collision and coalescence of water droplets. Droplets carried upwards by ascending currents fall relative to the rising air. In doing so, the larger droplets fall relative to their smaller neighbours.

Very small droplets are unable to collide with one another. However, when the radius of one droplet exceeds about 18 μ, collisions may occur. For larger drops the collision efficiency increases considerably.

Collisions and coalescence are necessary for the development of precipitation from warm clouds. The temperature of these clouds is above 0°C and they consist entirely of water droplets.

Water droplets also occur in many clouds whose temperatures are below 0°C. The supercooled droplets of these cold clouds may grow by colliding and coalescing with one another in a similar way.

Some cold clouds may also contain ice crystals. Let us now consider the various processes by which these may increase in size.

12.4 The formation of ice crystals

In Section 11.1 we discussed ice-forming nuclei, which are often referred to as ice nuclei. An ice nucleus is any particle which serves as a nucleus for the formation of ice crystals in the atmosphere.

These nuclei are often referred to as freezing nuclei. It is assumed that the initial development of an ice crystal takes place by the freezing of supercooled water about the nucleus.

When an ice crystal has been formed it may grow either by the direct change of water vapour to ice (i.e. deposition) or by the freezing of supercooled water droplets. In the next section we shall discuss how the first process takes place in mixed cold clouds. Later, we shall consider the formation of larger ice structures, which occur when ice crystals collide with supercooled water droplets.

12.5 The Bergeron process

A Swedish meteorologist, T. Bergeron, suggested a mechanism by which ice crystals may grow in mixed cold clouds. These are clouds composed of both ice crystals and supercooled water droplets.

In order to understand this process, you should refer again to Table 6.2. Notice that the saturation vapour pressure over ice is less than that over water at the same temperature.

In 1911 Wegener suggested that in mixed clouds the saturation vapour pressure adjusts itself to a value between the saturation values over ice and over water. Bergeron adopted this theory in 1933 to explain the growth of ice crystals in mixed cold clouds.

When an ice crystal exists in the presence of supercooled water droplets the situation is unstable. The water vapour is unsaturated with respect to water and so the water droplets evaporate.

At the same time, the water vapour is supersaturated with respect to ice. It therefore changes directly to ice and is deposited on the ice crystals. Thus, the ice crystals grow at the expense of the water droplets.

At temperatures only slightly below 0°C, the growth rate of ice crystals in mixed clouds differs little from that of the liquid water droplets. However, at temperatures below −10°C ice crystals grow more rapidly than the supercooled water droplets. The direct vapour transfer is most efficient at temperatures near −15°C.

It was originally assumed that almost every raindrop originated as an ice crystal in mixed cold clouds. For this to be the case, it would be necessary for all rainclouds to extend well above the 0°C level.

This is obviously not the case. For instance, in the tropics heavy showers develop from cumuliform clouds, whose tops are at temperatures above freezing point.

Actually, the Bergeron process is primarily concerned with the initial growth of ice crystals. Growth by direct deposition from water vapour may be rapid, while the ice crystal is small. However, the growth rate decreases as the crystal becomes larger.

Some type of collision process is necessary to produce large structures. Thus, in supercooled clouds, some collisions may lead to the freezing of water droplets on an ice crystal or snowflake. In other cases, snowflakes may form from the clustering together of ice crystals.

We have previously mentioned in Section 12.3 that collisions between droplets may lead to an increase in size and growth of water droplets. Let us now consider the growth of ice crystals by collision processes.

12.6 The growth of ice crystals by collision

Once an ice crystal has grown appreciably larger than the water droplets, it begins to fall relative to them. Collisions then become possible and growth of the ice crystal can proceed more rapidly.

Collisions with supercooled droplets may lead to the freezing of the water on to the surface of the ice crystal. This process is known as accretion.

The accretion of supercooled water may lead to the formation of either rime or glaze. Rime is a white or milky deposit of ice. It is formed by a relatively slow process, in which individual water droplets are collected by the ice crystal.

The supercooled droplets freeze immediately on impact and air is trapped between the particles. This prevents the transmission of light and so rime is opaque. Snow pellets are formed by the accretion of rime to an ice crystal. The resulting structure is porous and of low density. Snow pellets were formerly referred to as 'soft hail'.

Glaze is a coating of smooth ice. It is generally transparent (i.e. objects can be seen through it). However, if some air is trapped within it, glaze may be translucent (i.e. objects cannot be seen clearly through it).

The accretion of glaze to an ice crystal may proceed at a rapid rate on collision with large supercooled droplets. Two stages can be distinguished. First, the latent heat released by the freezing of part of the supercooled liquid may prevent the remainder from freezing immediately. The ice therefore becomes covered with a thin layer of supercooled water.

During the second stage of growth, the layer of supercooled water freezes relatively slowly. A dense more or less transparent mass is then formed over the outside of the structure. Deposits of glaze are found on ice pellets and hailstones.

Snowflakes consist of aggregates of ice crystals. These may occur in an infinite variety of shapes and forms. The largest snowflakes occur when the temperature is slightly below freezing point.

Under these conditions, riming assists in the collision and aggregation of ice crystals to form snowflakes. Rimed particles have different fall speeds according to their size, thus making collisions possible. The presence of liquid on the surfaces of the particles may also help to cement them together.

If any of the various ice structures discussed in this section fall below the freezing level, melting may take place. On emerging from the cloud, the raindrops may be indistinguishable from those formed by the coalescence of water droplets. However, in cold weather or when large ice structures are formed, the precipitation may still reach the ground in the solid state.

12.7 **Summary of precipitation development**

In the preceding sections we discussed the growth of both water droplets and ice crystals in clouds. Clouds form and develop in the atmosphere almost entirely as the result of the expansion and consequent cooling of ascending moist air.

When the rising air is lifted above the condensation level the most active condensation nuclei grow into droplets. Further cooling below 0°C does not, however, necessarily lead to freezing of the droplets.

Supercooling may occur if ice-forming nuclei are not present in the air. These are different particles from condensation nuclei and are much less common. For this reason, ice crystals do not usually occur in appreciable numbers in cloud-tops until their temperature has fallen to about −20°C.

The initial growth of water droplets by condensation of water vapour is rapid, but the growth-rate decreases as the droplet becomes larger. Similarly, the growth of an ice crystal by the Bergeron process is rapid while it is small, but decreases as it increases in size.

Some type of collision process is necessary to produce liquid or solid particles of precipitation size. The larger water droplets and ice crystals fall relative to the smaller particles. If they are larger than a critical size, they collide with the particles in their path.

Water droplets may coalesce to form larger droplets. Ice crystals may grow by the accretion of supercooled liquid droplets which freeze on contact, or by clustering together to form snowflakes.

Eventually, the water droplets or ice structures reach a size at which the ascending air currents are no longer able to sustain them within the cloud. Collisions may take place between precipitation particles of different size and fall rate. This may lead to the growth of some particles. However, there is a tendency, as they descend below the cloud base, for the water droplets to evaporate and for the ice structures to melt or sublime.

If the particles are able to survive the evaporation they experience while falling through the unsaturated air below the clouds precipitation is said to occur. If they do not reach the ground, but hang for some distance below the cloud base, they are referred to as virga.

12.8 **Characteristics of precipitation forms**

Falling hydrometeors that eventually reach the ground are referred to as precipitation. These may leave the cloud either as water drops or as various ice structures.

For these hydrometeors to reach the ground in their original state the air below the cloud must not be too warm or too dry. Otherwise, the water drops may evaporate and the ice structures either melt or sublime.

Drizzle consists of fairly uniform precipitation, composed exclusively of fine drops of water very close to one another. By convention, the radii of the drizzle drops are less than 250 μ. Drizzle frequently falls from stratiform cloud, which is only a few hundred metres thick. It only reaches the ground if the upward currents are very slight.

Rain is usually composed of liquid water particles of larger size than drizzle. These raindrops have radii larger than 250 μ. However, precipitation in the form of widely scattered smaller drops is also referred to as rain.

Large raindrops are produced by clouds that are usually many kilometres in thickness. The highest rate of rainfall occurs when relatively large drops are produced by cumuliform clouds. These clouds may sometimes be 10 km or more thick, and strong vertical currents occur within them.

Snow is precipitation in the form of ice crystals. Most of these are branched and are sometimes star-shaped. Aggregates of ice crystals are called snowflakes.

Sometimes there is a fall of unbranched ice crystals in the form of needles, columns or plates. These are known as ice prisms. They are often so tiny that they seem to be suspended in the air.

Snow grains consist of very small white and opaque grains of ice. These grains are fairly flat or elongated and their diameter is generally less than 1 mm. They neither shatter nor bounce when they hit a hard surface. Snow grains usually fall in very small quantities, mostly from stratus cloud or fog.

Snow pellets consist of white and opaque grains of ice. These grains are spherical or sometimes conical in shape, and they are about 2 to 5 mm in diameter.

Snow pellets are formed when the accretion of supercooled water to an ice crystal or snowflake is in the form of rime. They differ from snow grains in that they are larger, crisp and easily crushed. They rebound when they fall on a hard surface and often break up.

Ice pellets are a form of precipitation consisting of transparent or translucent pellets of ice. They are spherical or irregular in shape and have a diameter of 5 mm or less. They usually bounce on hitting hard ground and make a sound on impact.

WMO has subdivided ice pellets into two main types:

(a) Frozen raindrops or largely melted and refrozen snowflakes
(b) Pellets of snow encased in a thin layer of ice (formerly called 'small hail').

Hail is precipitation in the form of small balls or pieces of ice. An individual unit of hail is called a hailstone. It may have a diameter ranging from 5 to 50 mm, or sometimes more. Smaller particles of similar origin are referred to as ice pellets.

Hailstones may sometimes be composed entirely of transparent ice. However, they usually consist of a series of transparent layers alternating with translucent layers.

It is assumed that hail forms when some type of ice structure falls into supercooled water. A collision and accretion then occur. If it is a large droplet, it may cover the ice particle first and later freeze to form a transparent layer. By contrast, smaller supercooled drops may freeze on contact, trapping air and producing a translucent layer.

These processes would occur above the freezing level. Alternate transparent and translucent layers could arise from the hailstones traversing different parts of the cloud. Violent updrafts and downdrafts that occur in cumulonimbus clouds would be conducive to this development.

Some hailstones show evidence of layers being formed by alternate freezing and melting. It has been suggested that oscillations up and down through the freezing level may have occurred in these cases. Recent investigations have, however, indicated that the foliated structure could be produced merely by descent through an updraft.

Cumulonimbus clouds are favourable for the formation of hail. These clouds are characterised by strong updrafts, large liquid water contents, large cloud-drop sizes and great vertical height.

Hail therefore frequently occurs during thunderstorms. However, falls of hail take place within small sections of thunderstorms. Hence, an observer on the ground may not notice any hail.

Generally hail has to fall through many kilometres of air where the temperature is above 0°C. Therefore the hailstones may melt before reaching the ground. This perhaps accounts for the fact that hail is rarely observed at low levels in equatorial regions.

In studying precipitation we have been concerned with hydrometeors that reach the ground. As these liquid and solid water particles fall through the air the visibility of the atmosphere is affected. Lithometeors, such as dust, smoke and other solid particles, also make it more difficult to distinguish distant objects. Let us now investigate how visibility is measured and what factors cause it to vary.

CHAPTER 13

VISIBILITY

We often find it difficult to see distant objects by day and distant lights at night. This is because small particles are present in the air. These may include hydrometeors (fog, mist, rain, snow, etc.) or lithometeors (dust, smoke, etc.).

Some of the light from a distant object may be absorbed by these particles. However, scattering is the process which is mainly responsible for reducing the visibility. Under hazy or dusty conditions most of the light from a distant object may be scattered by particles in the air before it can reach the eye.

In this chapter we shall consider how visibility is defined in meteorology and some of the ways in which it can be affected.

13.1 **Meteorological visibility**

In meteorology, visibility refers to the transparency of the atmosphere in relation to human vision. It is expressed as a certain distance. For convenience, a given value for the visibility refers to the same condition of the atmosphere by night as by day.

Meteorological personnel are concerned with horizontal visibility. They report the greatest visibility prevailing over half or more of the horizon. If a significantly different visibility occurs in one direction, it is usually mentioned separately. It should also be realised that visibility from an aircraft is different from the horizontal visibility, which meteorologists measure near the earth's surface.

Everyone is aware of the fact that a dark building on the skyline may be far more easily seen than a sheep in a less distant bare paddock. This raises the questions of contrast against the background and the angular size of an object. Let us investigate how these factors are taken into consideration.

13.2 **Definition of day visibility**

In daylight, meteorological visibility is defined as the greatest distance at which a black object of suitable dimensions can be seen and recognised against the horizon sky. It should subtend an angle at the observer's eye of at least 0.5° horizontally and vertically. At the same time, however, the object should not be so large in the horizontal direction that it subtends an angle of more than 5°.

It is of interest to note how an angle of 0.5° may be estimated by punching a hole in a card. If the diameter of the hole is 7.5 mm, it will subtend an angle of about 0.5° when held at arm's length. The visibility object viewed through such an aperture should completely fill it.

To a certain extent the identification of an object depends on the observer's familiarity with the surroundings. It should be noted that the definition requires the observer to recognise the shape of the object against the horizon sky. It is not, however, necessary for him to be able to identify specific details.

13.3 **Night visibility**

The above definition cannot be used at night. The meteorological visibility is then defined as the greatest distance at which the black object of suitable dimensions could be seen and recognised, if the general illumination were raised to the normal daylight level.

In practice, the most suitable objects for determining visibility at night are unfocused lights of moderate intensity at known distances. The silhouettes of hills and mountains against the sky may also be used.

Sighting lights at night is more difficult than observing objects by day against the horizontal sky. The ability to distinguish faint lights depends on the general illumination around the observer. After coming out of a bright light, it may take up to half an hour for one's eyes to become completely adjusted to the surroundings. For this reason, visibility should be the last of the outdoor observations to be made at night.

In general darkness, a faint light can be discerned more readily when the observer is not looking directly at the light and when he is moving his point of vision.

The colour of the light being viewed also has an effect on the observer's ability to discern it. If the general illumination is weak, it is more difficult to identify a violet light than a red one.

For practical purposes, the relationship between the brightness of a given light and the daylight visibility may be expressed in one of the following ways:

(a) The greatest distance at which a light of luminous intensity equal to 100 candela can be seen
(b) The luminous intensity of a light that is just visible at a specified distance.

Tables based on these relationships have been established for use in night visibility observations.

13.4 **Visibility meters**

Meters have been devised to give a measurement of visibility, but they are not completely accurate. There is therefore no advantage in using an instrument for day-time measurements if a suitable series of objects is available for direct observation.

By contrast, visibility meters are useful for night observations. They are also of value if no distant objects are available for sighting as, for example, on board ship.

13.5 **Factors affecting visibility**

The main causes of reduced meteorological visibility are as follows:

(a) Precipitation
(b) Fog and mist
(c) Wind-blown spray from the sea
(d) Oils
(e) Smoke
(f) Dust and sand
(g) Salt.

13.6 **Precipitation effects**

In the case of precipitation, the obscurity may be caused by either water droplets or ice particles. Sometimes both types of hydrometeors occur together.

The visibility in rain depends on both the size of the droplets and their number in a given volume of air. Light rain has little effect, but moderate rain usually reduces the visibility to 3 to 10 km. In heavy rain, however, the visibility may be as low as 50 to 500 m.

In drizzle, the visibility depends on the intensity and may range from about 0.5 to 3 km. If fog droplets are also present the visibility may be less than 0.5 km.

Snow has more effect than rain. In moderate snow, the visibility commonly falls below 1 km. In heavy snow, it may range from about 200 m to less than 50 m.

The effect of wind may be important. Blowing snow is formed when snow is blown off the ground. This is likely to take place if the snow is dry and powdery. Serious reductions in visibility may therefore occur when the temperatures are low and in high latitudes.

13.7 **Fog and mist**

A fog usually consists of water droplets, but ice crystals may be present in some circumstances. At high latitudes 'ice fogs' may occur if the temperature is below −20°C, provided that the wind is light and other conditions are favourable.

Over high ground a fog can be regarded as a cloud on the ground. It may be one of the usual cloud types that requires adiabatic ascent for its formation.

In general, however, condensation is usually associated with a cold underlying surface. Two distinct types occur in this way:

(a) Radiation fog—caused by cooling of the ground by nocturnal radiation
(b) Advection fog—caused by the transport of relatively warm air over a colder surface.

Radiation fogs occur on cloudless nights. The ground is cooled by radiation, and conduction lowers the temperature of the air in its immediate vicinity. However, air is a poor conductor and the cooling may be limited in depth to a few centimetres. Dew or frost may then form on the colder ground, removing water vapour from the air.

Any turbulence will, however, cause mixing of the air. The cooling may then be spread through a greater depth. If turbulence is sufficient, stratus cloud may develop (see Section 11.3).

If the wind is lighter and less turbulence occurs, fog may be formed. The conditions for the formation of radiation fog are critical, some places being more prone to fog due to local effects. In general, the conditions necessary for the formation of radiation fog are a sufficiently high dew-point, sufficient cooling during the night and slight turbulence.

Advection fogs develop when air moves over a cooler land or sea surface, the surface temperature being below the dew-point of the moving air. Over the land, cooling by radiation frequently intensifies the effects of advection, when warm moist air arrives from the sea. This is described as an advection-radiation fog.

Advection is also responsible for the development of steam fogs. In this case, however, it is the cold air which is advected. As it passes over a warm water surface, water vapour evaporates into the air from the water. The effect is similar to that produced by hot water evaporating from a bathtub in winter.

Steam fogs do not involve any cooling of the air. In fact, there will be some heating. They depend for their formation on the addition of water vapour to the cold unsaturated air. These fogs are confined to water surfaces near a source of cold air, such as frozen land, or icefloes in polar regions.

The various types of fogs discussed above could be described as air-mass fogs. They depend on cooling taking place within an extensive and more or less uniform air mass.

By contrast, a frontal fog arises from the interaction of two air masses. This may occur in either of two ways. One type develops when cloud extends down to the surface during the passage of the front. It is generally located over hills rather than over low ground. The other type of frontal fog occurs when the air becomes saturated by evaporation from rain. This increases the dew-point of the air near the earth's surface. In particular, these conditions are likely to develop in the cold air ahead of a warm front. Pre-frontal fogs associated with warm fronts may be very widespread.

Fogs are very effective in reducing visibility. All colours of visible light are affected uniformly. The physical processes which produce fog also occur in the development of mists. When the water droplets in the air reduce the visibility below 1 km, meteorologists describe the phenomenon as a fog, instead of a mist.

Most fogs are dispersed either by turbulence or by warming. Although slight turbulence is necessary for the development of fogs, vertical mixing by increased turbulence may lead to their dispersal. Warmer and drier air is mixed with the fog and the fog particles evaporate.

Solar radiation is absorbed by the ground (and slightly by the fog). The ground then heats the adjacent air and the fog particles begin to evaporate.

Remember, however, that the upper surface of the fog is similar to the upper surface of a cloud. It therefore reflects some of the incoming short-wave solar radiation and reduces the amount reaching the ground. The warming of the ground is therefore much less than on a cloudless fog-free day. Fog by its presence therefore tends to preserve itself from dispersal by solar warming.

13.8 **Wind-blown spray**

As the wind speed increases over the sea, the crests of the waves become higher, and eventually spray begins to form from the breaking waves. With stronger winds, streaks of foam are blown through the air.

Once the wind speed reaches that of a strong gale (41 to 47 kn) the spray begins to affect the surface visibility. The effect increases with wind speed, and when it reaches hurricane force (>63 kn) the visibility may be seriously reduced.

The deterioration in visibility produced by wind-blown spray can similarly affect coasts with on-shore winds.

13.9 **Oil particles in the atmosphere**

The visibility in some cities may be somewhat reduced by the presence of oil particles in the air. The escape of oil fumes from motor vehicles is the main source of these particles. In many cities a great deal of attention is being given to devising methods for reducing the air pollution produced by oil particles entering the air from industrial sources and from motor vehicles.

13.10 **Reduction of visibility by smoke**

A thick haze is often produced by smoke from industrial and domestic fires. The larger particles tend to settle, but much of the pollution in the form of finely divided particles remains suspended in the air. These small particles are often comparable in size to water droplets in a mist or fog.

Most materials which are burnt produce minute particles of carbon. The carbon of smoke produces the black haze that is characteristic of large industrial cities. Coal and wood, when burnt, produce various tarry substances, which with the carbon particles produce soot.

Many fuels contain small amounts of sulphur, which forms sulphur dioxide (SO_2) gas on burning. The action of sunlight on sulphur dioxide leads to the formation of sulphur trioxide (SO_3). This gas is very hygroscopic and water droplets form on it. Sulphuric acid (H_2SO_4) is produced in this way, and, in addition to reducing the visibility, presents a health hazard to persons living in industrial areas.

If an inversion is present, it checks rising currents and the smoke is confined to the lower layers of the atmosphere. In light winds and humid conditions fogs may also be present. The combined effect of smoke and fog often reduces the visibility very seriously. The term smog is given to this phenomenon. In addition to the inconvenience caused to road and rail traffic, it may present a serious problem for aviation and shipping.

Apart from its effects on visibility, smog is often a health hazard in many cities. Some of the products of combustion may be toxic.

13.11 **Reduction of visibility by dust and sand**

Dust or sand may be raised from the ground by the wind and carried upwards. The height to which the particles rise depends on their size and the prevailing conditions.

When the visibility is reduced below 1 km, the phenomenon may be referred to either as a sandstorm or as a duststorm. Sandstorms consist of relatively coarse sand which is too heavy to be lifted far. The sand particles rarely extend to a height of more than 20 or 30 m, and they are not carried appreciable distances from their source.

Duststorms consist of minute particles of fine dust, and may on occasions be distributed to heights of several kilometres above the earth's surface. For dust to be raised and to remain in the air, a number of conditions must be satisfied.

First, the land surface must be dry and dusty. Second, the wind should reach at least moderate speeds in order to disturb the dust. Third, the air must be unstable, if extensive vertical motion is to occur. In stable conditions turbulence induced by the wind is damped out and the dust only rises a few metres.

Duststorms may be prolonged and widespread, or else occur in patches in association with minor squalls of wind. In some cases, there are few or no clouds to impede the solar radiation. The ground is strongly heated and steep lapse rates are established near the ground.

In other situations, dust is raised when large cumulus and cumulonimbus clouds are present. Strong lapse rates are obviously in existence and the dust may even be carried to cloud level. Rain may later remove the dust from the air.

A more widespread type of duststorm occurs with strong winds of long fetch over desert. With unstable conditions such storms may persist for many hours. Visibility may be reduced to a few hundred metres, or in extreme cases to a few metres. Sometimes they extend into the night, but they are then generally less severe, owing to the reduced lapse rate.

The fine dust carried upwards by various types of duststorm gradually spreads through the atmosphere. Many of the particles are too small to fall by gravity at any measurable speed. As a result, air of desert origin retains its haziness for a long period of time. The final clearance by rain or snow may sometimes occur at places many thousands of kilometres from the dust source.

13.12 **The effects of salt particles**

Sea spray often evaporates after being blown into the atmosphere. Each droplet then leaves behind a particle of salt to become a condensation nucleus at a later time. Sea salt particles are hygroscopic, and water will sometimes condense on them when the relative humidity is as low as 70%. For the same concentration of particles, sea salt is much more effective than smoke in reducing visibility. The haze produced by sea salt has a whitish appearance.

The various factors which effect the visibility are, in turn, influenced by the motion of the atmosphere. In the next chapter we shall study the characteristics of various types of local winds.

CHAPTER 14

LOCAL WINDS

The nature of the earth's surface varies considerably. Oceans, deserts, snowfields, forests, lakes, grasslands, cities, etc. cover various parts of its surface. Over land areas, its elevation also differs from place to place as one crosses hills, valleys, mountains, etc.

In any locality, the nature of the atmospheric flow is influenced by the characteristics of the surface over which the air is flowing and by the varying elevation of the earth's surface. In this chapter we shall study some of these effects.

14.1 Sea breezes

Near the coast an on-shore wind frequently sets in during the late morning, rises to a maximum in the early afternoon and then dies away in the evening. The strength of the breeze is greater on warm days, but may be weaker under cloudy conditions. It is called a sea breeze.

The basic cause of the air flow is the different rate of heating of the land and sea surfaces when solar radiation falls on them. You will recall from Section 3.6 that, during the day-time the temperature of the sea does not rise as rapidly as that of a land surface. As a result, in the lower layers the air above the land surface becomes warmer than that over the sea.

Figure 14.1 Pressure differences arising from heating differences

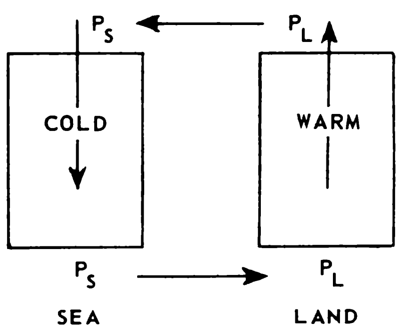


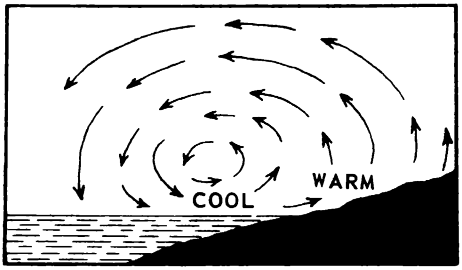
Figure 14.1 shows the effect on two air columns of equal height; one above the sea and the other over the land. The warm air above the land expands and rises. Some of the air enters the region above the top of the column.

The pressure (P_L) at this height then becomes greater than the pressure (P_S) over the sea at the same level. As a result, the air aloft tends to move towards the top of the cold column.

At sea level, the pressure (P_S) over the sea is higher than that (P_L) over the land due to the transfer of air aloft. It therefore tends to move from the sea to the land, i.e. a sea breeze develops. The circulation is completed as the cold air above the sea descends to replace the air moving towards the land.

The result of these various effects is shown in Figure 14.2. At latitudes greater than about 20°, the Coriolis force becomes strong enough to considerably affect the direction of the sea breeze as the circulation develops.

Figure 14.2 A sea breeze

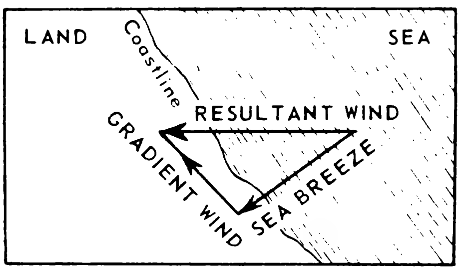


In the tropics, the contrasts in land and sea temperatures are quite marked. There is also a greater tendency towards the development of instability over the heated land. Sea breezes therefore tend to be stronger in these regions. If the air over the land is moist and unstable a thunderstorm may even develop after the onset of a sea breeze.

In some synoptic situations, the gradient wind may be in the opposite direction to that of the sea breeze. It will therefore delay its development, and in some cases prevent it from reaching the land. By contrast, if the gradient wind is approximately in the direction of the sea breeze, the resultant wind speed will be increased.

On the other hand, the resultant wind may sometimes take up a direction somewhere between that of the gradient wind and the sea breeze. This is illustrated in Figure 14.3, where the lengths of the arrows are proportional to the wind speeds.

Figure 14.3 Effect of sea breeze on gradient wind



As the temperature differences become greater in the early afternoon, the local pressure gradient between the sea and the land becomes steeper. The strength of the sea breeze component then increases. As a result, the Coriolis force is increased and the sea breeze effect may become directed more nearly parallel to the coast.

In the vicinity of large lakes, a similar process leads to the development of a lake breeze. These are, of course, on a smaller scale than sea breezes.

By contrast, a monsoon circulation is the result of heating differences produced on a large scale. They have not of course local winds, but develop between the ocean and an entire continent. The monsoons of India and other regions are produced in this way.

14.2 **Land breezes**

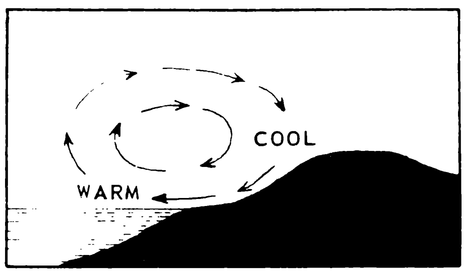
In coastal regions at night, land breezes may develop. These are directed in the lower layers from the land to the sea. Radiation cooling from the land surface takes place much more rapidly than from the adjacent ocean.

Eventually, the temperature of the land falls below that of the sea. The air in the lower layers of the atmosphere then cools more rapidly than that over the sea. In doing so, it contracts and descends.

The pressure aloft above the land then becomes lower than that at the same level over the sea. As a result, the air tends to move from the sea towards the land at these higher levels.

At sea level the situation is reversed. The pressure over the sea is lower than that over the land due to the transfer of air aloft. Consequently, the air in the lower layers tends to move from the land towards the sea, i.e. a land freeze develops. The complete circulation is shown in Figure 14.4.

Figure 14.4 A land breeze



In general, land breezes are not as strong as sea breezes. The heating differences are smaller and so the local pressure gradient is weaker. They are, once again, developed more markedly in tropical regions, where they may sometimes force moist unstable air to rise, leading to thunderstorms off the coast towards dawn.

14.3 **Katabatic winds**

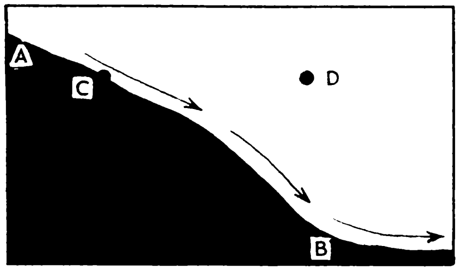
On cloudless nights, air often begins to flow down the slopes of mountains and hills. This downward flow becomes particularly evident as the air moves down the bottom of river valleys that lead to lower levels.

This type of air flow is called a katabatic wind. The name is derived from a Greek word meaning 'go down'. It develops at night, when the land surface loses heat by radiation.

Air which is cooled by contact with the cold land surface becomes denser than the surrounding air. The force of gravity on it is relatively greater and it is pulled down towards lower levels.

Figure 14.5 shows how a katabatic wind flow develops. AB represents the slope of a hill. The point C is located on the hill and D is a point at the same level in the free air.

Figure 14.5 A katabatic wind



Cooling by radiation on a cloudless night lowers the temperature of the ground surface at C. The air in contact with the land is then cooled by conduction. The air near C becomes denser than that in the free air near D, and a gravitational flow of air takes place down the slope.

As the air descends, it will be compressed at the higher pressures prevailing at lower levels. If this were the only effect, the air would tend to become warmer, and the flow would stop. On the other hand, the air continues to lose heat by contact with the cold land surface. This counteracts the adiabatic warming and so the katabatic flow is maintained.

In general, katabatic winds are rather light. In some situations, however, they may attain a considerable speed if the slope is steep and smooth. This is particularly the case if the surface is covered by snow or ice.

When highlands are close to the coast the katabatic wind may reinforce the land breeze at night. This may lead to rather strong off-shore winds.

14.4 **Anabatic winds**

Anabatic winds are produced by the opposite process to that causing a katabatic wind. An anabatic wind is a gentle upward flow on hill slopes on a fine warm day. The name is derived from a Greek word meaning 'go up'.

On a warm cloudless day, the ground slope becomes warmed by solar radiation and reaches a higher temperature than the air. However, air near the slope makes contact with it, and so becomes warmer than the free air at the same level.

The warmed air tends to become unstable and rises, being replaced by the surrounding air, which is cooler and denser. As it moves up the slope it expands because of the lower pressures aloft. Adiabatic cooling would tend to stop the ascent, if it were not counteracted by the continued heating of the air by contact with the warm slope.

In general, anabatic winds tend to be rather weak. The pressure gradients developed by the heating differences on a warm sunny day may be large. However, the air has to be forced uphill against the force of gravity. This reduces the speed of the flow up the slope.

14.5 **Föhn winds**

In Europe a warm dry southerly wind frequently flows down the northern slopes of the Alps. The local name for this wind is the föhn.

The physical processes which produce this wind also occur in other parts of the world. Meteorologists have therefore adopted this word as a general name for similar warm dry winds that develop in other mountain areas.

On meeting a mountain barrier, air is forced to rise and adiabatic cooling takes place. If the moisture content of the air is sufficiently high, the water vapour may condense to form water droplets. In some cases, ice crystals may be formed if the temperatures are sufficiently low and freezing nuclei are present.

During the cloud formation latent heat is released and this partly counteracts the adiabatic cooling of the rising air. It then cools at a slower rate—the saturated adiabatic lapse rate. The release of this latent heat is one of the essential processes that later lead to the development of föhn winds.

Precipitation in the form of rain or snow may fall from the rising air within the orographic cloud on the windward slope of the mountain. The precipitation therefore reduces the moisture content of the air, which continues to proceed over the mountain. This process is also essential for the subsequent development of a föhn wind.

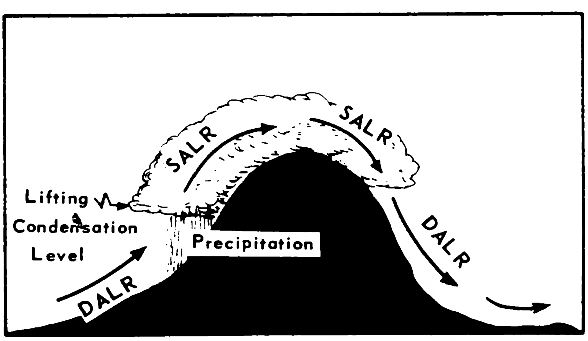
During the flow of air up the mountain, two important processes therefore take place. Latent heat from cloud formation supplies heat energy to the rising air. Loss of water droplets or ice crystals by precipitation makes the air drier than it was before it made the ascent.

Let us now study the motion down the slope. On the lee side the descending air warms due to adiabatic compression. Some cloud droplets evaporate and cool the air, partly counteracting the adiabatic warming. The air therefore warms at a lower rate —the saturated adiabatic lapse rate.

But the descending air is drier than it was before it was forced to rise up the mountain slope. The remaining cloud particles therefore evaporate after the air has descended a short distance down the slope. The cloud base is therefore higher on the lee side than it is on the windward side.

As it descends below cloud level, it now warms at the dry adiabatic lapse rate. The temperature rises rapidly as it travels from the high cloud base to the bottom of the slope.

Figure 14.6 The föhn wind



On arrival at the bottom its temperature is now higher than it was before it commenced its journey over the slope. In the greater depth of cloud on the windward side more latent heat was released than was later absorbed during the brief evaporation period on the descent.

The air is also drier, because some of the moisture was converted to precipitation. The föhn winds that reach the lower slopes on the lee side of mountains are therefore warm dry winds.

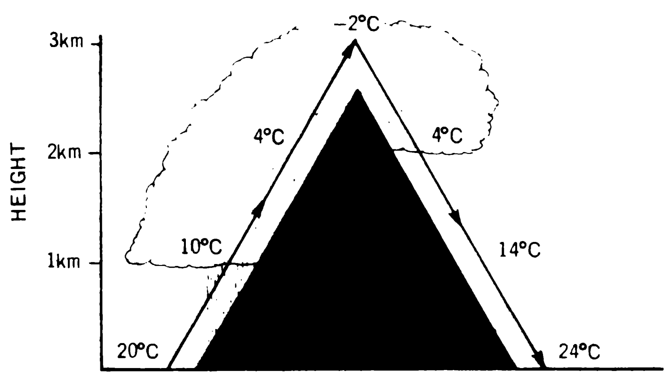
Figure 14.6 shows how the föhn wind develops. Notice the low cloud base on the windward side of the mountain. Remember that precipitation must occur in order to dry the air.

14.6 **Föhn wind temperature changes**

In order to illustrate the temperature changes that might occur during the development of a föhn wind, let us assume that the air rises to an elevation of 3 km. Suppose the cloud base is 1 km above the ground on the windward side. Assume that precipitation occurs on the ascent, and so the cloud base is higher on the lee side. Let the cloud base be 1 km higher than it is on the windward side.

Suppose the initial air temperature = 20°C, the DALR = 10°C/km and SALR = 6°C/km. Figure 14.7 then illustrates the temperature changes which occur. Notice how the temperature of the air has been increased by its passage over the mountain.

Figure 14.7 Temperature changes during föhn wind development



Not all mountain winds produce the föhn effect. If the air is too dry for the formation of clouds, it rises and cools at the dry adiabatic lapse rate throughout. The warming on the descent equals the cooling on the ascent and the temperature returns to the same value.

If cloud forms, but no precipitation occurs, the cloud base may be the same on each side of the mountain. Suppose no precipitation had occurred on the ascent in the example shown in Figure 14.7. Check for yourself the temperature changes that would occur on the descent, if no mixing takes place with the surrounding air. Notice that the variations in temperature would be the reverse of those on the ascent.

The final temperature would therefore be the same as the initial temperature—it would not be a föhn wind.

Local winds of various types are responsible for many of the characteristic features of the weather in localities where the surface of the earth is not uniform. Their effects are often quite striking, especially if they provide a trigger action for other atmospheric processes. In the next chapter we shall study some of the more violent phenomena which occur in the atmosphere and investigate how they may be produced.

CHAPTER 15

SEVERE LOCAL STORMS

Some weather systems are often too small to be detected on a standard synoptic weather chart, yet they can produce weather effects of great local importance. The local winds that we discussed in the previous chapter are of this type. More violent phenomena, such as thunderstorms and tornadoes, are also included in this category.

In this chapter we shall be concerned with the characteristics and physical processes associated with some of the more common types of severe local storms.

15.1 **Significance of scale in meteorology**

A great variety of weather phenomena occur, ranging in size from small eddies to thunderstorms, depressions and larger features which occur on a hemispheric or global scale.

Micrometeorology is concerned with micro-scale weather processes. The word 'micro' is derived from a Greek word meaning 'small'. Micrometeorology is primarily concerned with processes, such as turbulence and evaporation, which take place in the layer next to the ground.

Weather phenomena such as sea breezes and thunderstorms are studied in mesometeorology. The prefix 'meso' is derived from the Greek word *mesos* meaning 'middle'. It is therefore concerned with weather phenomena of medium size, and these are often described as mesoscale features of the atmospheric circulation.

Still larger are the depressions (lows) and anticyclones (highs) that appear on synoptic weather maps. Synoptic meteorology is concerned with these types of weather systems, which occur on the synoptic scale.

Even larger features can be detected on hemispheric weather maps. Large waves and major circulation features occur on a macroscale. The study of these phenomena is known as macrometeorology. The prefix is derived from the Greek word *macros*, meaning 'large'.

In this chapter we shall be studying some mesoscale phenomena which produce severe weather conditions. Because they are often difficult to detect on standard synoptic charts, they are often referred to as local storms. This does not mean that their effects are unimportant. In fact, one of the most violent of all storms is the tornado, yet its devastating effects are confined to a relatively small area.

15.2 **Thunderstorms**

In Chapter 9 we discussed the various types of non-astronomical meteors which are studied in meteorology. An electrometeor is defined as a visible or audible manifestation of atmospheric electricity.

A thunderstorm is therefore an electrometeor. The World Meteorological Organization defines it as one or more sudden electrical discharges, manifested by a flash of light (lightning) and a sharp or rumbling sound (thunder).

The thunder is the noise of the lightning discharge. However, the speed of light is very much greater than that of sound. As a result, the sound of the thunder may be noticeably delayed if the lightning occurs at a considerable distance from the observer.

High winds, torrential rain, thunder and lightning provide evidence of the enormous amount of energy which is expended during a violent thunderstorm. This energy is largely derived from the latent heat released during the condensation of water vapour. Part of this heat is converted into kinetic energy and accounts for the violent winds that accompany the thunderstorm.

Thunderstorms occur in convective clouds and are usually associated with precipitation, which reaches the ground in the form of a shower of rain, snow pellets, ice pellets or hail.

The formation of hail was discussed in Section 12.8. Hail and lightning are frequently associated with one another, but the hail often melts before reaching the ground.

A lightning flash is a large-scale example of an electric spark. A spark or discharge occurs when the electrical potential difference between two points separated by air reaches a very high value. This value depends on the conductivity of the air and the distance between the points.

Electrical potential differences develop in clouds when positive and negative electrical charges are separated from one another. There is still a great deal of uncertainty as to the nature of the actual mechanisms that produce these conditions.

It is known, however, that the outermost surface of water droplets is largely made up of negative charges, while just beneath this layer is one of positive charges. The frictional forces that occur during violent storms could remove the outer layer from these droplets and so separate the charges. Charge separation may also occur when water freezes, or when ice crystals melt.

Investigations by sounding balloons and by other means reveal the main features of the charge separation that occurs in thunderstorm clouds, namely

(a) positively charged area at the top of the cloud
(b) a concentration of negatively charged particles in the central region of the cloud.

Below the negatively charged region there is often a second, but more localised area of positive electricity.

When the electrical potential difference between the cloud and the ground or between two clouds or between parts of a cloud exceeds the breakdown potential, an electrical discharge (lightning) occurs.

15.3 The life history of a thunderstorm cell

It is often possible to recognise individual turrets bulging upward in a growing part of a convection cloud. At other times, connected masses or lines of thunderstorms extend over a horizontal distance of more than 50 km.

It is sometimes possible to associate a thunderstorm with a unit of convection known as a cell. The diameter of the thunderstorm cell may be about 10 km. An isolated cell may form from several growing cumulus clouds. On other occasions, active turrets appear above a broad cloud mass.

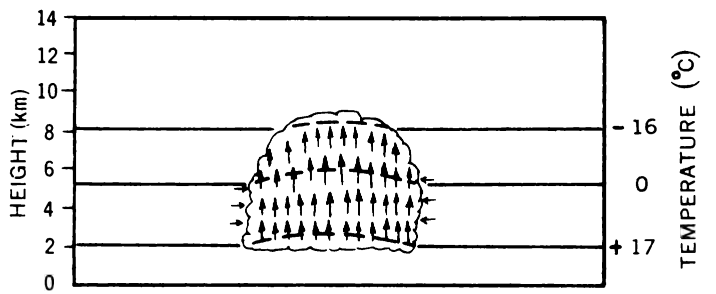
In general, there is a tendency for adjacent cells to join together. However, they can usually be distinguished through their separate precipitation patterns or radar echoes. Aircraft often pass through a less turbulent region that lies in the connecting part between cells.

The life history of a thunderstorm cell can be divided into three stages, based on the speed and direction of the vertical currents:

(a) The growing stage
(b) The mature stage
(c) The dying stage.

In the growing stage strong vertical up-currents occur throughout the cloud. Although pilots report rain or snow within the cloud, these appear to be suspended by the updraft, because no precipitation reaches the ground at this stage. Figure 15.1 shows this situation, which lasts 10 to 15 minutes.

Figure 15.1 The growing storm



The mature stage begins when water droplets or ice particles distinctly fall out of the bottom of the cloud. Except in arid regions, they usually reach the ground as precipitation. The size and the concentration of the droplets or ice particles are now too great for them to be supported by the updrafts in the cloud.

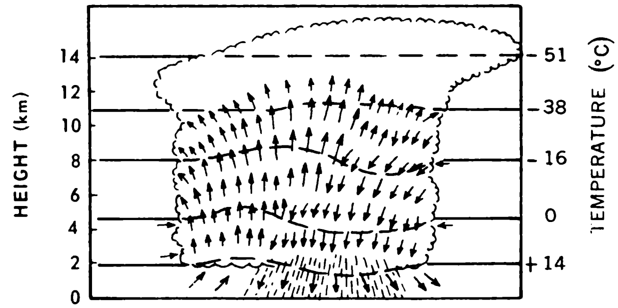
The frictional drag exerted by the falling hydrometeors helps to change the updraft into a downdraft in some parts of the storm. The updraft, however, still continues and often reaches its greatest strength in the upper parts of the cloud in the early mature stage.

The downdraft is usually not so strong and is most pronounced in the lower part of the cloud. The descending air is forced to spread out laterally near the earth's surface, often creating severe wind squalls. In the downdraft the temperatures are lower than those of the surrounding air. Figure 15.2 illustrates this effect.

The mature stage of the thunderstorm cell may therefore be accompanied by violent effects near the earth's surface. These include strong downdrafts of cold air, squalls, torrential rain and often hail. This stage lasts 15 to 30 minutes.

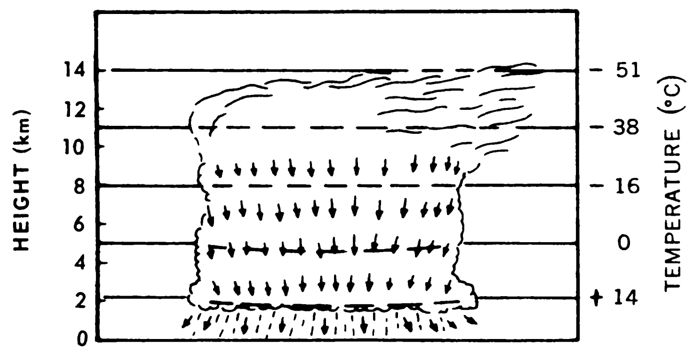
In the dying stage the updraft disappears entirely. The downdraft spreads over the entire cell and condensation can no longer occur. Once the supply of falling water droplets and ice particles ceases, the downdraft also weakens. This is indicated in Figure 15.3.

Figure 15.2 The mature storm



While the downdraft and the rain persist the entire cell is colder than the surrounding air. By the time they cease the temperature within the cell is restored to about the same value as that of the surroundings. Dissipation is then complete and only a few stratified clouds remain. At the surface all signs of the thunderstorm and the downdrafts have disappeared. This stage lasts about 30 minutes.

Figure 15.3 The dying storm



15.4 Types of thunderstorms

Thunderstorms may occur in a variety of synoptic situations. The initial requirements for their development are as follows:

(a) The presence of moist air through a considerable depth of the atmosphere
(b) An unstable atmosphere for saturated air extending to great heights
(c) A strong lifting mechanism to force air aloft to considerable heights.

Thunderstorms may be classified according to the immediate meteorological processes involved in their formation. Six types are usually distinguished:

(a) Heat
(b) Nocturnal equatorial
(c) Cold stream
(d) Orographic
(e) Convergence
(f) Frontal

Of these, the first five may develop within an air mass and are known as air mass thunderstorms.

Heat thunderstorms are characteristic of humid summer weather over continental areas. In those parts of the world affected by monsoons (discussed in

Chapter 19), heat thunderstorms occur sporadically prior to the onset of the monsoon.

Nocturnal equatorial storms occur over some tropical oceans and are most frequent near sunrise. We have seen that the diurnal variation of temperature over a sea surface is very small. On the other hand, some cooling of the tropical atmosphere can occur at night by long-wave radiation from water vapour and tops of existing cumulus clouds. This steepens the environmental lapse rate sufficiently to allow cumulonimbus clouds to develop.

Cold stream thunderstorms may occur where cold air flows fairly rapidly over warmer seas. Heat is added to the lower troposphere, thus increasing the temperature lapse rate within the air mass. Such thunderstorms may occur around the southern coast of Australia during winter and early spring in southerly airstreams, but they are not usually as violent as the other types.

Air mass thunderstorms may also occur where the lifting mechanism is provided by orographic ascent, or by widespread ascent caused by convergence within a trough of low pressure. These factors were discussed in Chapter 11.

Thunderstorms may also develop as the result of the interaction of two air masses. These are known as frontal thunderstorms. The lifting occurs either by cold air undercutting warm moist air or by warm moist air overriding colder air beneath.

Sometimes a combination of more than one process provides the trigger action for the lifting of the moist unstable air. The lifting mechanism itself also tends to steepen the lapse rate, and so stimulate thunderstorm development.

15.5 Detection of thunderstorms

Thunderstorms are mesoscale weather phenomena. They are often too small to be detected on a standard synoptic chart due to the distance between the reporting stations.

Sometimes a close network of meteorological reports are obtained from stations within a confined area, such as near a large airport. Mesoscale analysis may then detect the presence of a local thunderstorm.

In recent years more and more use has been made of electronic equipment to locate thunderstorms. One method involves the location of the electrical discharges of thunderstorms. Lightning often produces static on radio sets, and this fact can be put to use to locate a thunderstorm.

When lightning occurs, some of the energy is emitted in the form of low-frequency radio waves. These tend to follow the curvature of the earth with very little reduction in strength, and can be detected thousands of kilometres away.

The lightning discharges are known as atmospherics or sferics. Equipment used to locate the electrical discharges, and so the thunderstorm, is known as sferics equipment. Several widely spread stations, connected by telephone or radio, all obtain the direction of a single flash, and its location can be determined by triangulation. Sferics observations are useful in detecting thunderstorms, which may be located up to two or three thousand kilometres away over oceans or uninhabited areas.

The electronic detection of thunderstorms may also be accomplished by the use of radar equipment. A special type of radio transmitter emits radio waves of a certain wavelength. Water droplets and ice crystals reflect, scatter and absorb the radio waves falling on them.

If the particles are greater than a certain size some of the energy may travel back along the same path and be intercepted by the receiving aerial. The reflected signals may then be observed as radar echoes on a display screen.

Rain drops and snow crystals are large enough to produce radar echoes. Precipitation patterns can be detected several hundred kilometres away by this means. Thunderstorms produce radar echoes of characteristic shape and development.

15.6 **Tornadoes**

Tornadoes are the most violent disturbances of the atmosphere, yet they are too small to be detected on a standard synoptic chart. They may range in diameter from values of less than a hundred metres to perhaps a kilometre or so.

These roaring, whirling storms may produce windspeeds of over 500 km/h as they move over the earth's surface. They tend to travel with a speed of about 50 km/h, and the length of their path is usually only a few kilometres. However, on occasions, some have apparently remained active for distances of a hundred kilometres or more.

A tornado always results from excessive instability in the atmosphere, and therefore steep environment lapse rates. It is also closely associated with vigorous thunderstorm activity. In fact, the name is derived from the Spanish word *tronada*, meaning thunderstorm.

The tornado first appears as a funnel-shaped cloud, extending down from the base of a cumulonimbus cloud. After it touches the ground, the funnel creates havoc, destroying buildings and sucking debris and dust up into the air. Motor cars, animals and heavy objects may be drawn aloft and dumped several hundred metres away.

Apart from the strong winds, the destruction of buildings is partly accomplished by an explosive effect. The pressure may fall by over fifty millibars in less than a minute. The large pressure differences between the inside of a closed building and the atmosphere outside then leads to an explosion, which bursts the walls and the ceiling outwards.

On the average, nearly 200 tornadoes occur each year in the USA. The Mississippi valley is the greatest tornado region in the world. Tornadoes occur, however, on all continents and cause great destruction to property in populated areas.

15.7 **Waterspouts**

Waterspouts occur over the sea and can cause considerable damage to any vessel in their path. They are of two types.

One type of waterspout develops downwards from a cumulonimbus cloud and is simply a tornado occurring over the water. The other builds upwards from the surface of the water and is not directly associated with a cloud.

Both of these are called waterspouts because water is drawn aloft, just as tornadoes suck up dust and debris over the land. The type which is not associated with a cloud is much less violent. It corresponds in activity with the 'dust devils' that occur over strongly heated deserts.

We have seen that local storms can be associated with violent effects and cause great havoc. Special methods must be used for their detection because the presence of mesoscale features is not evident on synoptic charts normally used at meteorological stations. These charts are, however, invaluable for locating the presence of synoptic scale systems such as depressions (lows) and anticyclones (highs). These systems are associated with large-scale weather phenomena and will be studied in later chapters.

CHAPTER 16

AIR MASSES AND FRONTS

When air remains over a large uniform surface for several days or weeks, it tends to acquire the characteristics of the surface below. If the air is colder than the underlying surface, it will be warmed and the heat from the surface may be transferred upwards through a layer several kilometres thick. Similarly, air which remains over an ocean surface will gradually acquire moisture.

Thus, the temperature and moisture content of the air tend to approach a state of balance or equilibrium with that of the surface below. Just how close it will eventually come to reaching this state depends on the circumstances. Obviously, the time spent in the region will be important.

In this chapter we shall first consider situations in which a considerable thickness of air has reached a temperature and moisture distribution characteristic of the underlying surface. In these circumstances it will tend to retain these properties for some time after it leaves the source region. Later, we shall study what happens when it meets air which has spent some time in a different region.

16.1 Definition of an air mass

When air possesses similar characteristics over a large area it is called an air mass. Level for level, its temperature and moisture content are approximately the same over large horizontal distances.

16.2 Source regions

In order that a mass of air may assume uniform characteristics, it needs to remain more or less stagnant for a period of many days. To be described as an air mass, it must cover a large area where the earth's surface itself is reasonably uniform. Such a surface is called an air mass source region.

Stagnant air is most likely to be associated with large stationary or slow-moving anticyclones (highs). In the vicinity of the centres of these high pressure areas the pressure gradients are weak, and calm or light winds prevail over large areas of the earth's surface.

It will be shown later (see Figure 20.1) that these conditions are most likely to be found in the vicinity of the sub-tropical high pressure belt and of the polar high pressure region of each hemisphere.

Air may also remain for considerable periods in other localities. For instance, large high pressure areas frequently occur over continents during winter.

16.3 Classification of air masses

A simplified picture of the general circulation of the atmosphere sometimes shows a single front in each hemisphere—the polar front. In general, the polar front of each hemisphere circles the earth in middle latitudes. However, it may meander equatorwards or polewards, forming a series of waves around each hemisphere.

The warm air on the equatorward side of a polar front is described as being a 'tropical air mass'. It may, of course, often include air from sub-tropical as well as tropical regions.

On the poleward side of a polar front the air is cold and it is called a 'polar air mass'. It does not necessarily originate over polar regions, but may form over subpolar areas as well.

Only two types of air masses are included in this simple picture—a warm tropical air mass and a cold polar air mass. This simple description of the atmosphere is approximately true in the middle and upper troposphere. However, the picture is much more complicated in the lower troposphere.

Complications arise in the lower troposphere for two main reasons. First, the circulation is much more complex at low levels. Many temporary or transitional types of air masses develop and these form detached fronts that are often shortlived.

Second, the continents and oceans impart different properties to the overlying atmosphere. Contrasting air masses are therefore created. These differences are more marked in the lower troposphere than at higher levels.

One way of classifying air masses is to refer to their source regions. This practice is often useful in regions near the air mass source.

However, in dealing with large sections of the earth, it is impossible to use the source name for very long. The air eventually changes its characteristics as it moves to a different region. Thus the source name refers only to the recent history of the air mass.

When referring to the lower troposphere some meteorologists do, however, use a general classification, based primarily on the latitude of the source region. In order of increasing latitude they use the following terms:

(a) Equatorial air
(b) Tropical air
(c) Polar air
(d) Arctic (or antarctic) air.

The factor most affected by latitude is the temperature. The distinction between (a) and (b) is often difficult to determine, as sharp differences in temperature are rarely maintained for long periods in the warmer regions of the earth. However, arctic (or antarctic) air is extremely cold and dry (because it can hold very little moisture). It is therefore sometimes possible to distinguish a front between (c) and (d).

A secondary classification takes into consideration differences in moisture content between two air masses. Air which has its origin over the ocean has a high moisture content and is called a maritime air mass. By contrast, an air mass formed over a land surface is relatively dry. It is known as a continental air mass.

Radiosonde observations usually provide evidence of the characteristic properties that air masses acquire from their source regions. Air masses can also be identified by the contrasts which appear between them along the fronts on synoptic charts.

16.4 Air mass symbols

The simplest classification based on source regions merely takes into consideration the combined effects of temperature and moisture content. Tropical (T) and polar (P) air masses are assumed to be relatively warm and cold respectively. Maritime (m) air masses are considered moist and continental (c) air masses dry.

Four air masses can therefore be distinguished, and the following symbols are sometimes used to distinguish them:

Tropical maritime	— Tm
Tropical continental	— Tc
Polar maritime	— Pm
Polar continental	— Pc.

Air masses, however, undergo modification as they move from their source regions. Cold air passing over a warm surface will be heated from below. By contrast, warm air will lose heat in its lower layers, if it moves over a colder surface.

These temperature changes affect the environment lapse rate and so the stability of the atmosphere. If the air becomes unstable, water vapour may be carried to higher levels. On the other hand, the development of a temperature inversion may prevent the vertical transfer of moisture.

Pronounced effects therefore occur in air masses that are warmer or colder than the surface over which they are passing. This leads to a further sub-classification of air masses and the following symbols are used:

(a) Air colder than the surface over which it moves (K)
(b) Air warmer than the surface over which it moves (W).

These symbols may be combined with source region symbols in the following way:

KTm	WTm
KTc	WTc
KPm	WPm
KPc	WPc.

Thus, KPm is a polar maritime air mass which is colder than the surface over which it is moving. This could occur if it were travelling equatorwards over a warm surface.

By contrast, air from the same source (Pm) might move polewards over an extremely cold surface. If it is warmer than the underlying surface, the prefix W is used, i.e. WPm.

Notice that the symbols K and W do not refer to the actual air temperature, but to the difference between its temperature and that of the underlying surface. Thus, cold air may be designated by the symbol W, if it is warmer than the even colder surface over which it is moving.

Let us now investigate some of the changes that occur when air masses leave their source regions.

16.5 The modification of air masses

When moving over a surface warmer than itself, an air mass (K) is heated from below. Thermal instability then develops in the lower layers and spreads upwards. If the air originally contained inversions, these will be destroyed, and a uniform steep lapse rate will be established in the lower troposphere.

If the air mass (K) travels over water, its moisture content will increase. Convection will transfer the water vapour to higher levels, where condensation and cloud formation will take place. Cumuliform clouds develop—cumulus, large cumulus and finally cumulonimbus may appear in turn. Showers and even thunderstorms may occur.

If the air mass (K) moves over land, less moisture will be absorbed. The development of convective cloud may then be delayed, until the heating from below extends the instability to much greater altitudes.

By contrast, an air mass (W) moving over a surface colder than itself loses heat in its lower layers and becomes increasingly stable. This may completely prevent convection from occurring.

The cooling from below results in a cooled surface layer of air. The air above the inversion, however, is mainly unaffected except for the slow cooling caused by outgoing radiation. Eventually, the air near the earth's surface may be cooled below its dew-point, in which case fog or stratus may form. Poor visibility and perhaps drizzle may occur.

In general, it is assumed that the characteristics of air masses change slowly. For instance, polar air penetrating the tropics and remaining there must be slowly modified before it can be labelled tropical air.

16.6 General characteristics of fronts

It is impossible to detect a sharp interface or boundary between two air masses. Rather, there is a zone of transition in which the properties of one air mass gradually change to those of the other. It is strictly more correct to use the term frontal zone, but the word front is generally used in synoptic meteorology.

The frontal zone may be several kilometres across, but a line is a suitable marking on standard synoptic weather charts. When it is desired to emphasise the gradual nature of the change from one air mass to another, the term 'frontal zone' is then used.

In the simplest sense, a frontal zone separates air masses of different densities. On synoptic charts differences in the temperature, and often the humidity, can be detected in the vicinity of frontal zones. These differences are much more marked than the negligible changes which occur over considerable distances within the individual air masses.

16.7 Classification of fronts

In Section 11.6 we discussed frontal depressions. They include the polar front depressions that develop along the polar front separating tropical and polar air masses. Let us study this development in more detail.

For simplicity, assume that the polar front is a plane surface. It is not vertical, but slopes polewards from its section with the earth's surface.

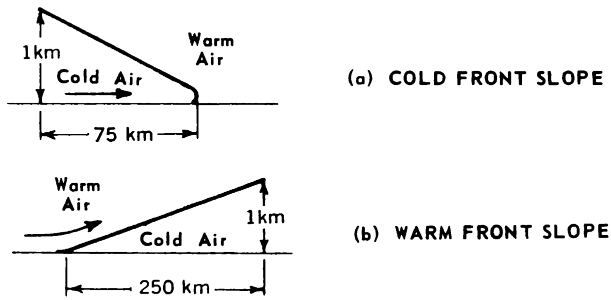
In some regions along the polar front, cold dense air advances equatorwards, causing warm air to be forced aloft over its sloping surface. This portion of the polar front is known as a cold front. Cold polar air is replacing warm tropical air.

In other regions along the front, warm air of lower density moves polewards, sliding over its sloping surface. This portion is called a warm front. Warm tropical air replaces cold polar air.

In general, the cold front is steeper than the warm front. On the average, a cold front rises about 1 km in a horizontal distance of about 75 km. In the case of a warm front the slope is about 1 in 250.

Figure 16.1 illustrates these features, but the vertical dimension is exaggerated for simplicity. Care should also be used in applying the average figures, as sometimes the slope of a cold front may be less than that of a warm front.

Figure 16.1 Characteristic slopes of fronts



16.8 Weather associated with idealised fronts

The weather associated with fronts depends on a number of factors, including the characteristics of the air masses and the way in which they interact with one another. Meteorologists have, however, developed idealised models of cold and warm fronts.

We shall discuss the weather associated with these idealised fronts, but it must be remembered that in practice many variations may occur. For instance, if the warm air is dry and stable, cloud development may be limited and precipitation may not occur.

16.9 The idealised warm front

The cloud and weather associated with an idealised warm front are shown in Figure 11.6.

If the warm air is moist, the approach of a warm front may first be heralded by the appearance of cirrus and cirrostratus cloud in a continually denser sheet. If the over-running air is unstable and turbulent, cirrocumulus may be seen (i.e. a 'mackerel sky').

As the front comes closer, the warm air becomes lower. Middle level clouds, such as altostratus and altocumulus, develop.

Rain or snow may first fall as the altostratus cloud reaches its greatest density. Sometimes, however, the hydrometeors evaporate before reaching the ground and virga can be seen below the main cloud deck.

The precipitation intensifies as nimbostratus cloud develops. Lower clouds are also often present in the cold air. Evaporation of raindrops and the surface rain water, accompanied by turbulence, leads to the development of these lower clouds.

The actual weather conditions associated with a warm front depend largely on the characteristics of the warm air before it is lifted. Furthermore, in view of the slow rate of ascent of the air up the relatively gradual sloping surface, much of the heavy rain can only be explained by the presence of strong convection within the warm air. Therefore, this air must itself be unstable.

16.10 The idealised cold front

Cold fronts frequently possess a steep slope. They then act more violently in producing clouds and precipitation, when cold air replaces warm moist air. The front may be characterised by cumulonimbus clouds, gusty turbulent winds, heavy rain and sometimes thunderstorms. When the cold front interacts with moist unstable air, a squall line accompanied by sudden showers and a vigorous wind shift may occur.

A steep cold front produces within a very short distance the same amount of lifting as occurs over a much broader zone in advance of a warm front. It is therefore accompanied by a much narrower band of cloud and precipitation than that which occurs with a warm front. Its effects tend to be brief, but violent. The cloud and weather associated with such a cold front are shown in Figure 11.7.

16.11 Extratropical depressions

A marked concentration of potential energy occurs in the vicinity of the polar front. Nature, however, has produced a mechanism which provides for the release of this energy. It is known as a wave depression or wave cyclone.

Because it occurs in regions outside the tropics, it is also called an extratropical depression. The prefix 'extra' is actually derived from the Latin word meaning 'outside'.

Extratropical depressions may sometimes develop in the absence of a front. However, those that do form along a front are associated with a wavelike twist on the front. We shall therefore consider these wave depressions in more detail.

16.12 The wave depression

During the First World War the Norwegians were cut off from weather reports from surrounding areas, especially the oceans. In order to overcome this deficiency a dense network of weather stations was established. As a result of preparing synoptic charts for this region a great deal of detailed information on extratropical depressions was obtained. A number of meteorologists working at the Geophysical Institute in Bergen, Norway were responsible for the early discoveries that led to the development of a model for the migratory depression that has formed the basis for the so-called 'polar-front' theory of cyclones.

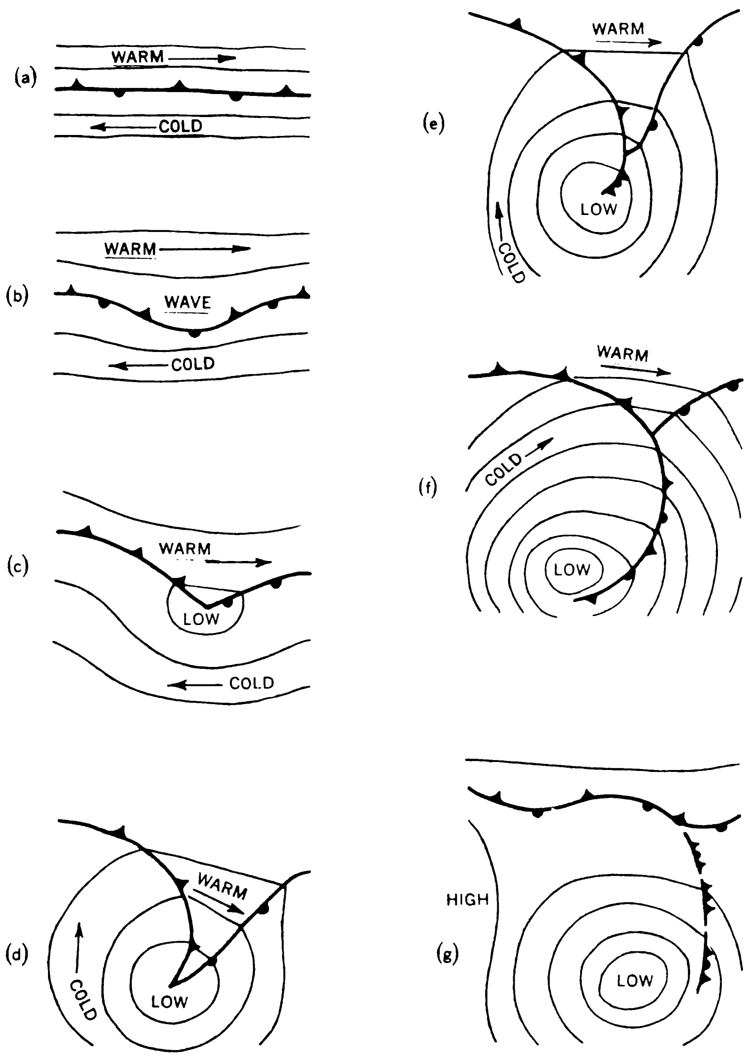
This theory recognises that extratropical depressions may develop on the polar front. Since then, it has been discovered that other fronts may develop, each of which has depression-forming possibilities. These depressions form at a wavelike twist on the front.

Some of these waves undergo little or no change and eventually die out. They are known as stable waves.

However, sometimes the amplitude of the wave increases, until great masses of polar or tropical air are carried away from their source regions. Eventually, these air masses become modified and mixed together. Waves of this type are called unstable waves.

Unstable waves grow in amplitude until they eventually appear to 'break' like ocean waves. Figure 16.2 shows how this may occur in the southern hemisphere.

Figure 16.2 Life history of a wave depression in the southern hemisphere



Diagrams (a) and (b) show the front separating warm tropical air from cold polar air with a wave beginning to form. As the wave develops, a low of increasing intensity forms around it, as in diagrams (c) to (f). In stage (g) the low gradually weakens.

16.13 Weather associated with a mature wave depression

Let us consider the weather that may be experienced when a depression approaches and passes a station, if it is at stage (d) of its development. Suppose we study what happens if the warm front, the warm sector and finally the cold front pass over the station in turn.

With the approach of the warm front there is a fall in pressure, which becomes more rapid as the front comes nearer. At the same time, there is an increase in cloudiness and precipitation, as well as in humidity. Generally the temperature is more or less constant or slowly rising until the surface front reaches the station.

There is then a sharp increase in temperature, the sharpness of which depends on the degree of contrast between the two air masses on either side of the front. With the passage of the front there is a decrease in cloudiness, or a complete clearing, as the station is now in the warm sector.

The region between the warm and cold fronts on the warmer side is occupied by the warmer air. This is the portion of the depression known as the warm sector.

In the warm sector the cloudiness depends on the general properties of the air mass that occupies that region. These include its temperature, moisture and lapse-rate conditions. The temperature, however, remains relatively high. The barometer is steady or shows a slight falling tendency at first. However, a more marked fall occurs as the cold front approaches.

With the approach of a steep well-developed cold front the winds in the warm sector begin to freshen and cumuliform clouds can be seen approaching, if the warm air is moist and unstable. The clouds lower as the front comes closer, and cumulonimbus cloud with increasing precipitation develops. With the actual passage of the cold front a vigorous squall occurs with a sharp change in wind direction as the cold air sweeps over the station. A marked rise in pressure occurs immediately after the front arrives.

Normally there is a fairly rapid clearing, if the slope of the cold front is steep. However, in mountainous or ocean areas, clearing showers may occur, if the air behind the front is moist and becomes unstable.

The above description refers to an idealised situation. It should be remembered that not all wave depressions follow this pattern. The cloud and precipitation characteristics associated with any particular depression depend on the temperature, moisture and lapse-rate conditions in the air masses.

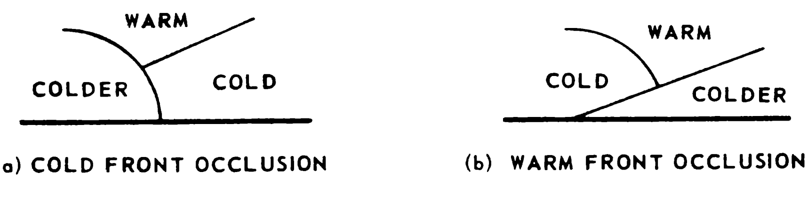
16.14 **Occluded fronts**

Diagrams (e) and (f) in Figure 16.2 illustrate what happens when the cold front moves faster than the warm front. As it overtakes the warm front, the warm sector is closed and a combined front forms. This process is called occlusion.

The front formed in this way is called an occluded front. Air originally in the surface warm sector is lifted aloft. Two different kinds of occluded fronts are possible, because the cold air behind the cold front has had a different life history from the cold air ahead of the warm front.

If the air behind the cold front is the colder of the two cold air masses, it will undercut the cold air under the warm front. A cold front occlusion is produced in this way. Diagram (a) of Figure 16.3 illustrates how this may occur.

Figure 16.3 Types of occluded fronts



In Figure 16.3 (b) a warm front occlusion is shown. In this case, the air under the warm front is the colder of the two cold air masses. As a result, the air behind the cold front rides over the very cold air mass.

16.15 **Weather associated with occluded fronts**

In both types of occlusions, a trough of warm air is pushed aloft from the warm sector. Clouds and precipitation may occur in this warm air as the cold air squeezes it upward. In many cases, interaction between the two cold surface air masses also produces clouds and precipitation in the lower levels accompanying the windshift line.

However, the actual type of weather associated with the occluded front depends upon its structure and action. Some of the characteristics of both cold and warm fronts may often be present, but the sequence of events is more complex than is the case before occlusion takes place.

As the occlusion process continues, the warm sector is displaced more and more aloft. The cyclone eventually becomes completely surrounded by cold air in the low levels. By the time this occurs, the air masses become either completely modified or mixed. The depression then decreases in intensity, until it dies out completely.

In this section we have been discussing an idealised model of a wave depression. In practice, more complex cloud and precipitation patterns frequently occur. However, the model provides a useful guide to the atmospheric processes that occur in extratropical wave depressions. This has been confirmed by experience in the past half century and by cloud photographs taken from earth satellites in recent years.

Meteorologists are interested in the development of many different types of weather systems. One of the most useful methods of studying these processes is to use synoptic weather charts. In the next chapter we shall discuss the characteristics of some of these charts.

CHAPTER 17

SYNOPTIC CHART ANALYSES

Meteorologists are interested in atmospheric processes that take place on a variety of scales. For instance, they may study the processes taking place within a small volume of air surrounding a growing crop. They then use special techniques to investigate this microscale problem.

Again, they may be concerned with processes which extend over a somewhat larger area. Mesoscale processes, such as thunderstorms and tornadoes, are often studied by making radar or sferics observations suitable for the detailed investigation of weather systems of this size.

Other meteorologists investigate the large anticyclones (highs), depressions (lows) and other systems that affect the weather over the continents and the great oceans of the earth. In order to study synoptic scale processes, it is necessary to obtain weather data from large parts of each hemisphere and through great depths of the atmosphere. In fact, meteorologists today feel that it is desirable to obtain observations from the atmosphere as a whole. The World Meteorological Organization has established the World Weather Watch for this purpose.

In order to study and interpret weather data on a synoptic scale, meteorological observers in the various parts of the world all make their observations at the same instant. The weather reports are then sent to a large meteorological centre, where the information is plotted on synoptic charts.

In this chapter we shall discuss some of the more important types of synoptic charts used by meteorologists. We shall also study some of the more common features appearing on analysed synoptic charts.

17.1 Types of synoptic charts

A variety of synoptic charts are used for studying both surface and upper air weather reports. Sometimes the observations of all stations are referred to the same level or altitude, and the data are then plotted on a constant-level chart. The most widely used constant-level chart is the mean sea level synoptic chart.

In studying upper air information, it is usually more convenient to obtain data from points in the atmosphere where the pressure has a particular value. For instance, meteorological observations may be obtained from points in the atmosphere where the pressure is equal to 500 mb. The data are then plotted on a constant-pressure chart. When studying atmospheric processes which occur near the middle of the troposphere, meteorologists often use the 500 mb synoptic chart.

After the meteorological data have been plotted on a synoptic chart, they are analysed graphically. In this way, a synoptic chart analysis is obtained. Characteristic patterns appear on analysed synoptic charts.

17.2 The mean sea level synoptic chart

The earliest synoptic charts were those which displayed the distribution of atmospheric pressure at mean sea level. In Section 5.9 we discussed how the station (level) pressure can be reduced to mean sea level. The mean sea level pressure at a given instant can be determined for a large number of places. These values can then be plotted on the MSL synoptic chart.

Lines joining places of equal pressure are called isobars. On many charts, isobars are drawn at 4 mb intervals, but in practice the interval used depends on the scale of the chart. After the MSL synoptic chart has been analysed a variety of isobaric patterns can be seen.

Weather observations are also plotted on the MSL synoptic chart. Temperatures, surface wind, present weather, cloud types, etc. are displayed on this chart. This information enables meteorologists to study weather developments and to locate fronts.

An analysed MSL synoptic chart displays both pressure and frontal systems. Let us investigate each of these in turn.

17.3 MSL isobaric patterns

Some isobaric patterns are very complex. However, they are all combinations of simpler basic systems that frequently occur on MSL synoptic charts. For simplicity, the pressure systems illustrated in the diagrams are shown as being fairly symmetrical.

Each basic pressure system has certain weather characteristics associated with it. These enable meteorologists to identify the systems. Reference will be made to these weather features.

17.4 Anticyclones and ridges

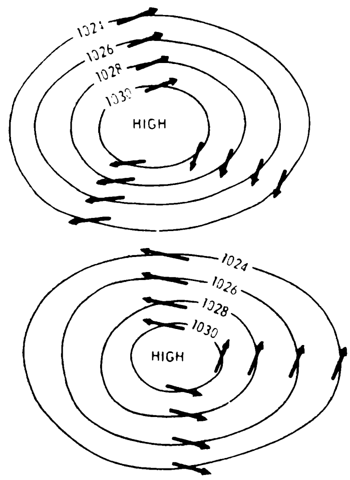
An area of relatively high atmospheric pressure is known as a high or anticyclone. The maximum pressure occurs at its centre, which is surrounded by one or more closed isobars. Fine weather and light winds frequently occur near the centre of a high.

In the northern hemisphere the circulation around a high pressure system is in a clockwise direction. By contrast, in the southern hemisphere it is anticlockwise. However, near the earth's surface frictional forces tend to cause the wind to blow slightly outwards across the isobars. Figure 17.1 shows this effect.

Figure 17.1 A high or anticyclone

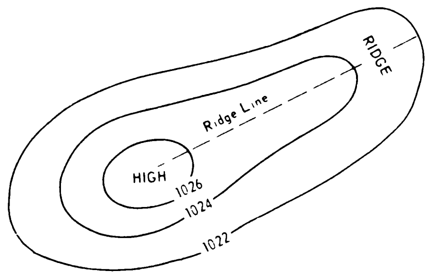
(a) NORTHERN HEMISPHERE

(b) SOUTHERN HEMISPHERE



A ridge is an elongated area of high pressure. It is indicated by rounded isobars extending outwards from an anticyclone and has associated with it a ridge line. The pressure at a point on the ridge line is higher than that at an adjacent point on either side of the line. Figure 17.2 illustrates the isobaric pattern of a ridge of high pressure.

Figure 17.2 A ridge of high pressure



17.5 Depressions and troughs

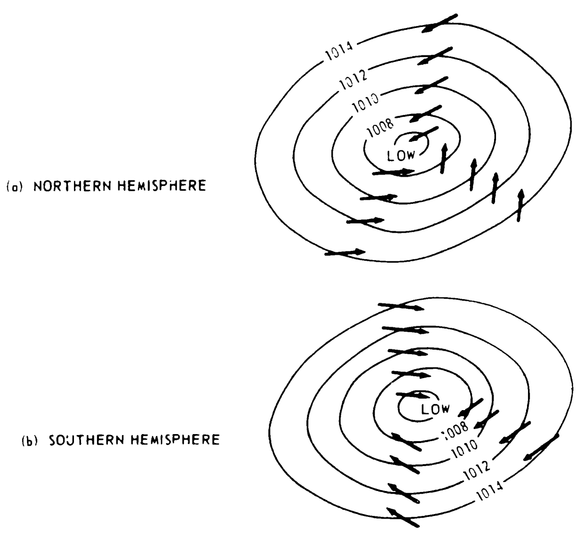
An area of relatively low pressure is referred to as a low or depression. The minimum pressure occurs at its centre and it is surrounded by one or more closed isobars.

A system of this type is sometimes called a cyclone. Unfortunately, in everyday conversation this name is often applied to special types of storms accompanied by strong winds. Although a low or depression is often associated with strong winds, this is not always the case.

In the northern hemisphere, the circulation around a low pressure system is in an anticlockwise direction. By contrast, it is clockwise in the southern hemisphere. Near the earth's surface, however, frictional forces tend to cause the wind to blow slightly inwards across the isobars.

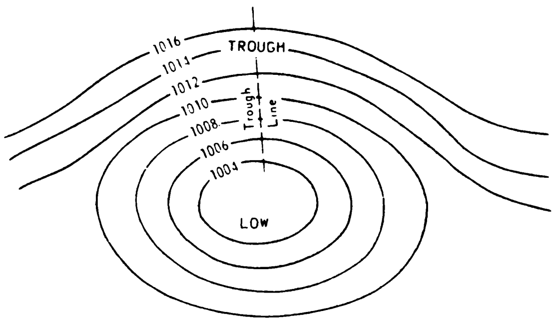
Figure 17.3 illustrates the surface circulations, which occur with low pressure systems in each hemisphere.

Figure 17.3 Low or depression (cyclone)



A trough of low pressure is indicated by isobars extending outwards from an area of low pressure. It has associated with it a trough line which is often indicated on a synoptic chart. The pressure at a point on a trough line is lower than that at an adjacent point on either side of the line. Figure 17.4 shows the isobaric pattern associated with a trough.

Figure 17.4 A trough of low pressure

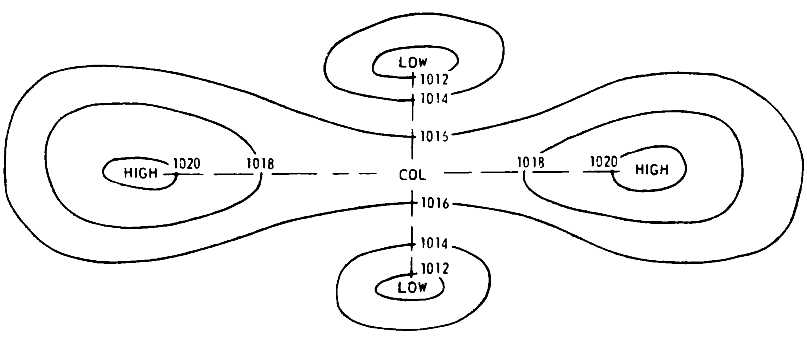


Low level convergence may be associated with the ascent of air in a depression or trough. If the air is moist and unstable, clouds and precipitation may occur.

17.6 Other MSL isobaric patterns

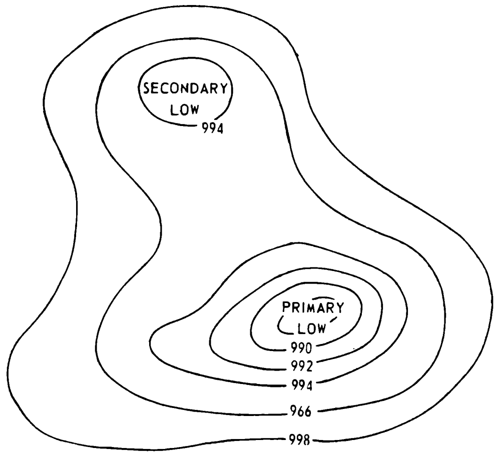
A col is a region between two highs and two lows. The centre of a col is located at the point of intersection of a trough line and a ridge line. Pressure gradients gradually change and reverse direction across a col. Near the centre of the region the pressure gradient is very weak, resulting in very light winds of variable direction. The isobaric pattern associated with a col is shown in Figure 17.5.

Figure 17.5 A col



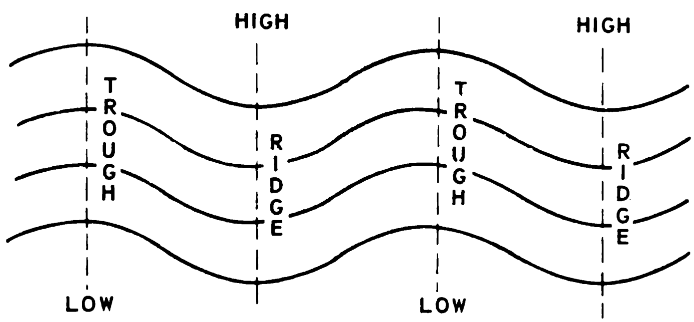
Another complex pattern occurs when a low pressure system forms near or in association with an existing primary low. It is called a secondary low (or depression) and is initially contained within the circulation of the original (primary) low. Later it may itself develop into a primary low. The arrangement of the isobars in Figure 17.6 shows the relationship between the two depressions.

Figure 17.6 Primary and secondary lows



Isobars sometimes form a wave pattern consisting of a series of troughs and ridges. These waves often extend for long distances approximately along the circles of latitude, and are then referred to as westerly waves or easterly waves. Figure 17.7 indicates an isobaric wave pattern.

Figure 17.7 An isobaric wave pattern



17.7 **MSL frontal systems**

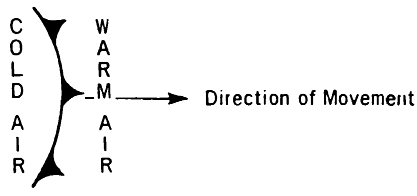
The transition zone between two air masses of different density is called a frontal zone or simply a front. Its sloping surface meets the ground in a region which may only be a few kilometres wide, if it is a sharp front. In the case of a diffuse front it may, however, be a hundred kilometres across. On the scale used for MSL synoptic charts, a front is generally represented by a line.

One of the most important factors affecting the density of the air is temperature. The air masses separated by the front are therefore usually of different temperature. Often differences in moisture content, stability, cloud formation and precipitation are also associated with a front.

A trough of low pressure is often located within a single air mass. In this case, it is not associated with a front. However, a trough is associated with a moving front. Two different air masses are then involved.

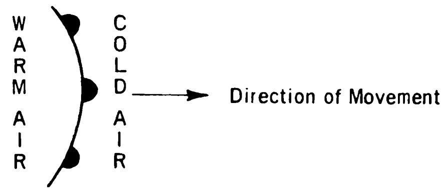
When a front moves towards the warmer air (i.e. cold air replaces warm air) it is called a cold front. On a synoptic chart a cold front is represented by a line faced with barbs pointing in the dirtcion of movement, as shown in Figure 17.8.

Figure 17.8 Representation of a cold front on an MSL synoptic chart



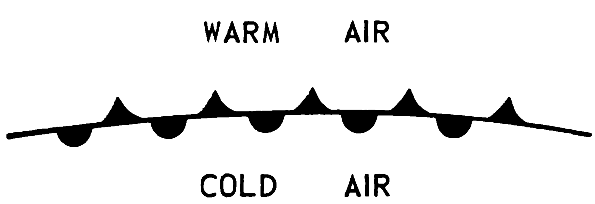
A warm front moves towards the colder air (i.e. warm air replaces cold air). It is represented by a line with semicircles pointing in the direction of movement. Figure 17.9 shows how it is indicated on an MSL synoptic chart.

Figure 17.9 Representation of a warm front on an MSL synoptic chart



If a front is not in motion, it is called a stationary front. In this case, the wind flow will be parallel to the frontal zone. A pressure trough does not occur in this case, and the front is parallel to the isobars. The barbs point from cold to warm air, while the semicircles point from warm to cold air (see Figure 17.10).

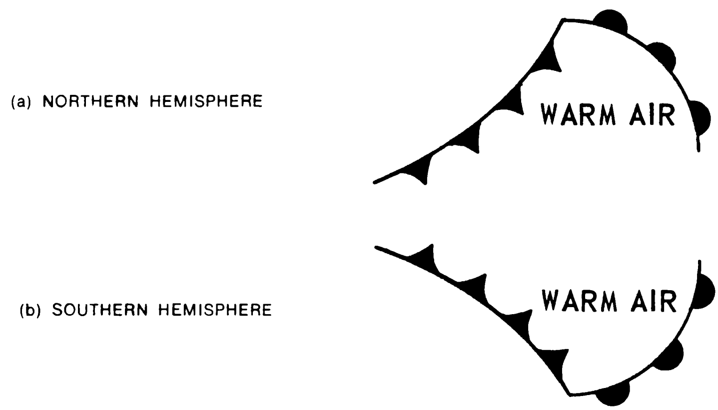
Figure 17.10 Representation of a stationary front on an MSL synoptic chart



In practice, slight movements of the front usually take place. It is then called a quasi-stationary front. The prefix is derived from the Latin word *quasi* meaning 'as if'.

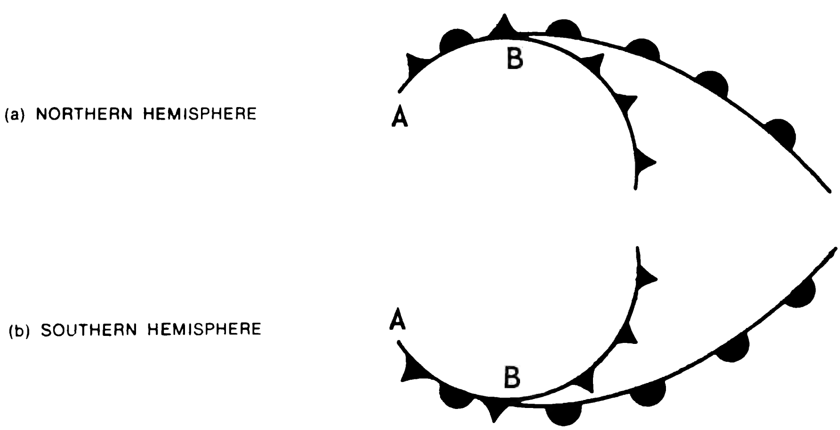
In Section 6.12, we mentioned that a frontal wave may sometimes develop leading to the formation of a type of extra-tropical depression, known as a wave depression or wave cyclone. Figure 17.11 shows how this frontal wave appears on a synoptic chart.

Figure 17.11 Representation of a frontal wave on an MSL synoptic chart



In the normal course of development of a wave depression, the cold front overtakes the warm front. Occlusion then takes place, as the warm sector is displaced more and more aloft. The occluded front is indicated by a line with alternate barbs and semicircles pointing in the direction of its motion. In Figure 17.12 the portion of the front marked AB is occluded.

Figure 17.12 Occlusion of a frontal wave



17.8 Constant-pressure synoptic charts

Radiosonde observations provide us with measurements of temperature and humidity for given pressure levels in the upper atmosphere. It is therefore convenient to draw constant-pressure charts at standard levels, such as 1000, 850, 700, 500, 300, 200 and 100 mb.

It is possible to draw lines on a constant-pressure chart, joining points that are at approximately the same altitude above mean sea level. These are known as contours. Various contour patterns, such as highs, lows, waves, troughs, etc. appear on upper air constant-pressure charts.

17.9 **Streamline analysis**

The horizontal wind is the key variable in any synoptic analysis. Wind analysis in regions outside the tropics is assisted by the fact that there is an approximate balance between the pressure gradient force and the Coriolis force.

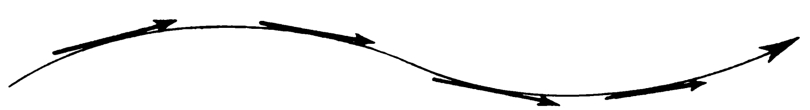
The circulation at the top of the friction layer (i.e. at about 1 km) can therefore be approximately determined from the MSL synoptic chart by observing the isobaric patterns. Upper wind circulations outside the tropics can similarly be determined from the contours of constant-pressure charts. Geostrophic wind scales can also be used for determining wind speeds in the vicinity of straight isobars and straight contours.

In low latitudes, however, the Coriolis force becomes relatively small. Winds then blow across the isobars or contours, and geostrophic wind scales are useless. At latitudes lower than about 20°, MSL isobaric charts and constant-pressure contour charts therefore lose a great deal of their value for wind analysis.

In order to overcome this difficulty, streamline analysis is frequently used in the tropics. The technique can be used at all levels.

It is possible to draw a line on a synoptic chart to represent the wind direction at a given instant. The only requirement is that the wind direction should be tangential to the line everywhere along its length. Such a line is called a directional streamline or simply a streamline. Figure 17.3 shows the relationship between the actual wind directions and a (directional) streamline.

Figure 17.13 Wind directions at a given instant indicated by a streamline



Unlike isobars and contour lines, the spacing between (directional) streamlines is only a matter of convenience. The spacing is not related to the wind speed.

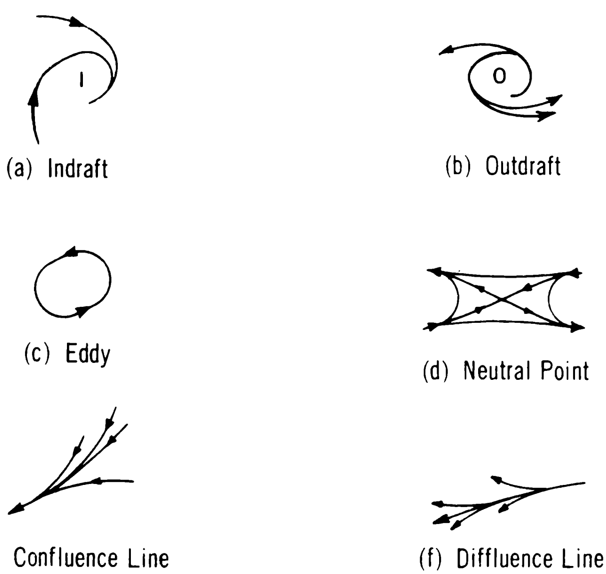
17.10 **Streamline patterns**

A number of characteristic streamline patterns are found on streamline synoptic charts. Some of the more important patterns are shown in Figure 17.14.

Outdrafts and indrafts correspond to the circulations associated respectively with highs and lows on MSL isobaric charts. A neutral point corresponds with the centre of a col.

Singular points are isolated points, where the wind appears to have all directions at once. The wind speed is actually zero at a singular point (i.e. calm conditions prevail). Singular points are located at the centres of indrafts, outdrafts and eddies. They also occur at neutral points.

Figure 17.14 Characteristic streamline patterns



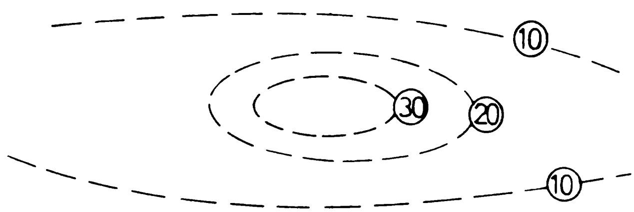
17.11 Isotach analysis

(Directional) streamlines provide no indication of the wind speed. Another technique is used to obtain a graphical display of the wind speed field. This is known as isotach analysis.

An isotach chart consists of lines joining points of equal wind speed. Each line is known as an isotach (Greek, 'equal speed'). The wind speeds are plotted (usually in knots) on the chart, and the isotachs are then drawn.

It is customary to display isotachs by broken lines. Figure 17.15 shows isotachs drawn for wind speeds of 10, 20 and 30 kn.

Figure 17.15 Isotachs (kn)



Sometimes an isotach analysis is conducted separately on an isotach chart. More frequently, streamlines and isotachs are displayed on the same chart. This is called a streamline-isotach chart.

In this chapter, we have dealt with common types of synoptic charts used in forecasting offices. Many other types of analyses are also required for predicting the weather, but a great deal of information can be obtained by studying the charts we have been discussing. In the next chapter, we shall consider some of the weather characteristics associated with the synoptic patterns found on meteorological charts.

CHAPTER 18

WEATHER ASSOCIATED WITH SYNOPTIC SYSTEMS

Synoptic systems include pressure systems, fronts and other patterns displayed on synoptic charts. The weather associated with these systems may be due to a variety of causes. These include such factors as the vertical motion induced by fronts, mountains, convergence, etc., and developments in the broad stream of air as it moves over the earth's surface.

A synoptic chart gives the overall picture for a particular instant. In extra-tropical regions isobaric charts indicate the approximate wind directions and speeds by the isobars and their spacing. Streamline-isotach charts provide similar information, by indicating the wind direction by streamlines and the wind speeds by isotachs.

18.1 The air circulation near the earth's surface

The MSL isobaric chart is used to determine the air circulation near the earth's surface in extra-tropical regions. However, frictional forces cause cross-isobar flow, and so the isobars are more representative of the circulation at the top of the friction layer (at about 1 km). If the winds are gradient winds (i.e. horizontal and at constant speed), then the isobars can be taken as approximating to the streamlines.

At latitudes less than 20° from the equator, gradient winds are rarely possible due to the weakness of the Coriolis effect. Isobars and streamlines are therefore markedly different in low latitudes.

The path of a parcel of air (i.e. its trajectory) does not necessarily lie along an isobar or streamline. If a pressure system remains stationary and the winds are gradient winds, the trajectory may be along an isobar. However, this situation rarely occurs.

Isobaric and streamline patterns gradually change with time. Synoptic systems move from one place to another, sometimes intensifying, at other times weakening. New systems appear and pass through their various stages of development.

Numerous weather situations are therefore possible. Let us now consider some general weather characteristics associated with a number of synoptic systems.

18.2 Anticyclones

This synoptic system occurs on MSL isobaric charts. Due to friction there is an outward cross-isobar flow of air near the earth's surface. This may be linked with the sinking of air (subsidence aloft), if the MSL pressure is rising due to upper air convergence. The environment lapse rate becomes more stable and an inversion develops.

The stability of the air is greatest near the centre of the anticyclone. If the air is dry, cloudless conditions may be associated with light and variable winds. Dew or frost may form at night.

If the air is moist, early morning mist or fog may occur in winter. During the daytime, a turbulence inversion may develop leading to overcast stratus or stratocumulus. Light drizzle may also occur.

In cities, the presence of an inversion in the central part of an anticyclone limits the vertical spreading of atmospheric pollution. Light winds or calm conditions also result in the pollution remaining near its source. Under an overcast cloud in badly polluted air, conditions are gloomy. Hence the term 'anticyclonic gloom'.

Away from the centres of anticyclones, conditions change as regions of lower pressure are approached. Winds become stronger and air is less stable. The actual weather depends largely on the nature of the underlying surface over which the air is moving. This affects the stability and moisture content of the air in the lower levels. Subsidence aloft may, however, still limit cloud development.

18.3 **Ridges**

The air in ridges is also relatively stable, owing to the effects of subsidence. The general weather characteristics associated with anticyclones therefore apply to ridges.

18.4 **Depressions**

Falling pressures are associated with rising air. This situation arises from the divergence of air at upper levels, and is linked with the inward cross-isobar flow which occurs near the earth's surface.

Frontal depressions are usually formed as the result of the development of waves on existing fronts. The formation and later development of a wave depression (i.e. wave cyclone) have been discussed earlier. The weather characteristics of cold, warm and occluded fronts have also been considered. It should be remembered, however, that the weather of each depression depends on the conditions prevailing at the time. Only the broad pattern of weather applies generally.

Some depressions do not develop from the formation of fronts. They are called non-frontal depressions. The weather occurring in various parts of the depression depends on the characteristics of the various streams. The weather conditions are, however, affected by the ascent of the air that occurs in the depression.

Some non-frontal depressions are formed as a result of intense inland heating. They are called heat lows. These depressions are often little more than wind circulations, since the low moisture content of the air may make it impossible for extensive cloud formation to occur.

Another type of non-frontal depression is, however, often accompanied by widespread rain. Strong divergence in the upper levels of the troposphere may lead to falling pressures at the earth's surface. A depression appears on the MSL synoptic chart and widespread slow vertical ascent takes place. This reduces the stability of the air and, if the air is moist, widespread cloud development and precipitation may occur.

Tropical depressions and cyclones are discussed in the next chapter. Tropical cyclones have a structure peculiar to them alone.

18.5 **Troughs**

Troughs are often associated with fronts. Frontal weather patterns have been discussed in earlier chapters.

Very often troughs appear with no front. These non-frontal troughs are, however, regions of relatively low pressure. Surface convergence and rising air therefore often lead to the development of cloudy conditions and bad weather.

18.6 **Cols**

A col is an area of very light winds. The weather depends on the characteristic of the particular air mass present. Diurnal effects often exert a strong influence.

18.7 **Stream weather**

A cold stream occurs when an air mass is colder at low levels than the surface over which is it moving. It is therefore heated from below and becomes less stable.

The depth through which the instability spreads depends on such factors as the amount of warming and the initial stability of the air. The characteristic cloud of a cold stream is cumulus, which often develops into cumulonimbus. Showers are common in cold streams.

Warm streams develop when air has spent some time over a warm area and then flows over a colder surface. It then loses heat to the underlying surface and becomes more stable. If the stream is moving over an ocean surface, the lower layers may become saturated. Fog or low stratus may then form.

18.8 **Streamline—isotach patterns**

Convergence in the lower troposphere with divergence aloft is associated with ascending air. Ascent may lead to cloud formation and precipitation.

By contrast, divergence near the earth's surface combined with convergence in the upper troposphere is associated with descending air. Subsidence leads to increased stability and reduces the development of clouds.

It is therefore important to be able to identify regions of convergence or divergence at any level. Usually it is necessary to consider both isotachs and streamlines.

Meteorologists use special techniques to determine whether convergence or divergence is occurring. Sometimes the weather itself can provide a useful guide to the wind field. For instance, bad weather associated with ascending air and cumuliform cloud implies surface convergence with divergence aloft.

Now that we have studied the type of synoptic analysis useful for low latitudes, it is appropriate to investigate the weather in the tropics in more detail. We shall therefore study tropical meteorology in the next chapter.

CHAPTER 19

TROPICAL METEOROLOGY

Geographically, the tropics lie between the Tropic of Cancer (23.5°N) and the Tropic of Capricorn (23.5°S). However, in meteorology it is frequently inconvenient to restrict the study of atmospheric processes to such precise limits.

For instance, a feature of the general circulation is the Hadley cell, which is located in low latitudes of each hemisphere with descending air reaching the earth's surface at about latitude 30°. It is often necessary to investigate the circulation in each cell as a whole, rather than to confine the study to the geographical tropics.

In many cases the weather in low latitudes cannot be studied in isolation. Interactions occur between portions of the atmosphere located in different latitudes and it is often desirable to consider the atmosphere as a whole. In this chapter, however, it will be convenient to direct most of our attention towards events occurring in areas that are of particular concern to tropical meteorologists.

19.1 The scope of tropical meteorology

During the year, each Hadley cell moves north and south in response to the apparent motion of the sun relative to the earth. The location of each cell is also influenced by the temperature differences, which arise from the uneven distribution of land and sea surfaces in each hemisphere. As a result, each cell changes its position in time in a variable way along the various meridians of the world.

The descending air of each Hadley cell reaches the earth's surface in the sub-tropical high pressure belt of each hemisphere ('the horse latitudes'). The centre of each belt of high pressure must therefore also move north and south about a mean position at about latitude 30°.

We shall therefore include in our study of tropical meteorology the region between the centres of the sub-tropical high pressure belts, i.e. the zone lying approximately between the 30th parallels of latitude.

Studies of the general circulation of the atmosphere indicate that the air reaching the earth's surface in the sub-tropical high pressure belts turns equatorwards to form the trade winds of each hemisphere. We shall therefore include the trade winds in this study of tropical meteorology.

19.2 The equatorial trough

The trade winds of each hemisphere move equatorwards from the sub-tropical high pressure belt towards a zone of relatively low pressure. This is known as the equatorial trough.

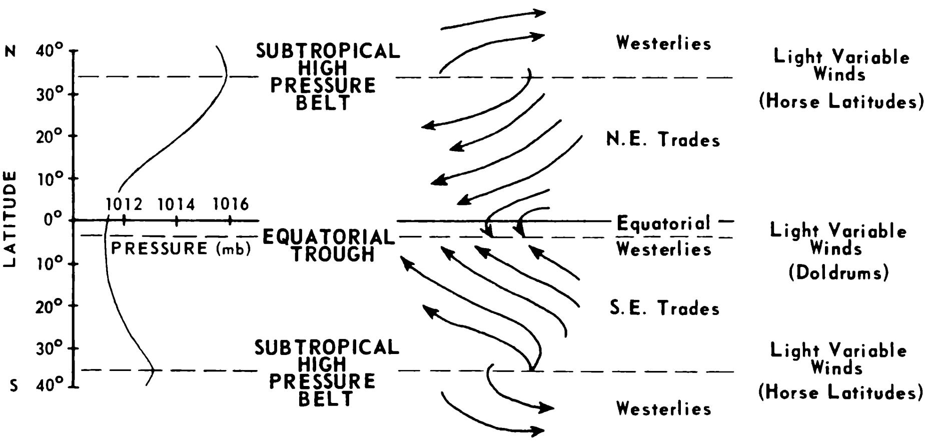
The mean position of the centre of the equatorial trough is not at the geographical equator, but at about latitude 5°N. This is sometimes called the meteorological equator.

19.3 The idealised average MSL circulation

Actually, the pressure and wind systems move north and south with the apparent motion of the sun relative to the earth. The two extreme positions occur in January and July.

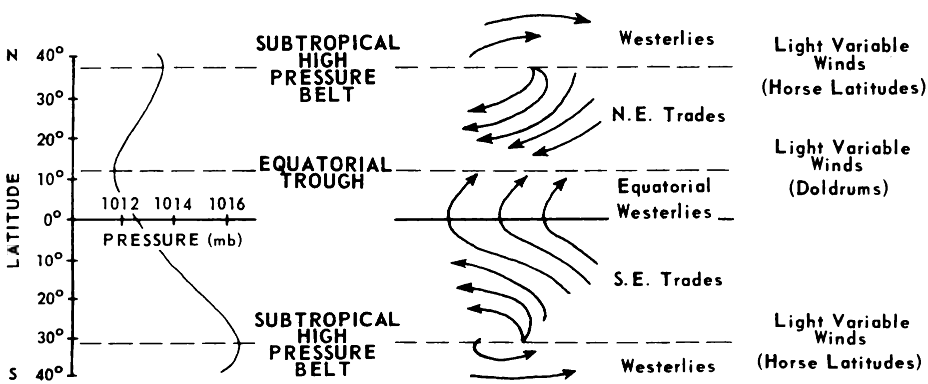
Figure 19.1 indicates the average MSL pressure distribution for January. An idealised picture of the average MSL circulation is also shown. At this time of the year, the centre of the equatorial trough is located at about latitude 5°S.

Figure 19.1 MSL pressure and streamlines in January



In Figure 19.2 the corresponding pressure and wind distributions for July are shown. The sun has now moved northwards and the equatorial trough is centred at about 12°N.

Figure 19.2 MSL pressure and streamlines in July



Approaching the equator, the Coriolis effect becomes very small. The wind is then strongly influenced by local effects and friction. As a result, there is a relatively large cross-isobar component towards low pressure (i.e. towards the equatorial trough).

After entering the other hemisphere, the Coriolis force increases with latitude, but the direction of the deflection is now reversed. The trade winds therefore tend to change to equatorial westerlies after crossing the equator on their way towards the centre of the equatorial trough. This effect can also be seen in Figures 19.1 and 19.2.

In some regions, however, easterlies may occur on both sides of the equatorial trough. In these areas the equatorial westerlies may be absent.

We have, however, been discussing the mean flow averaged around the earth for the January and July monthly periods respectively. In practice, variations in the position of the equatorial trough occur along different meridians.

Figure 19.3 Mean positions of the equatorial trough

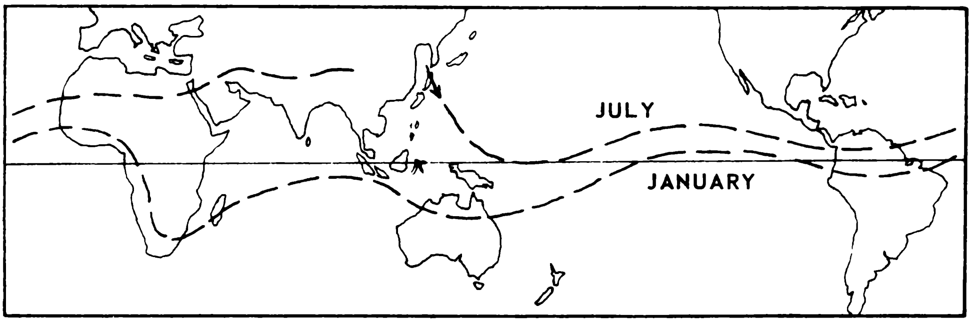


Figure 19.3 shows the mean positions of the equatorial trough along the various meridians. Notice that it passes over northern Australia and southern Africa in January. Later in the year it moves northwards and extends over southeast Asia and northern Africa in July. Larger movements occur in these regions than elsewhere, because heat lows develop over these continents, and so the average monthly MSL pressures are low.

The average monthly MSL pressures and the corresponding idealised wind patterns conceal the daily variations that occur in the tropics. Nevertheless, they provide a useful background for studies of tropical meteorology.

19.4 The trade winds

Most of the regions between the sub-tropical high pressure belts and the equatorial trough are occupied by the trade winds. In the northern hemisphere, the air moving equatorwards is deflected towards the right by the Coriolis force and the northeast trade winds develop. In a similar way, in the southern hemisphere a deflection towards the left produces the southeast trade winds.

The winds do not necessarily blow exactly from these directions. Apart from local effects due to variations in the underlying surface, differences in pressure gradients and the Coriolis force cause the wind direction and speed to vary from time to time and from place to place.

Nevertheless, the trade winds are generally noted for their persistence and steadiness. Over the ocean areas they are characterised by cumulus clouds with bases at about 1 km and tops at 2 km or so.

On coastlines facing the trade winds, the vertical cloud development is greater and there is a tendency to showers, especially if the wind speed increases. Regions on the lee side of mountain ranges may, however, sometimes be almost free of low clouds.

The restricted cloud development and the generally fine weather associated with the trade winds are linked with a trade wind inversion. Subsidence in the sub-tropical high pressure belt produces an inversion, which persists for part of the equatorward motion of the air as a trade wind. The trade wind inversion places a 'lid' on cloud development, especially over ocean areas.

However, as the air moves equatorwards and westwards towards the equatorial trough, the trade wind inversion begins to weaken and its base becomes higher. Vertical cloud development then increases as the instability extends to greater altitudes. Precipitation becomes heavier and more frequent in the vicinity of the equatorial trough.

19.5 **The monsoons**

The main wind systems are modified by the effects of land and sea surfaces. Because of the different thermal properties of soil and water, a land surface heats and cools more rapidly than does an ocean region.

Continents therefore tend to warm up quickly in summer and to cool rapidly in winter. By contrast, the seas maintain a more even temperature throughout the year. The effects of heating differences between land and sea surfaces were discussed in Sections 14.1 and 14.2. When large continental and ocean areas are involved, air circulations may develop on a grand scale.

In summer, air begins to flow from the oceans towards the hot continents. At the same time, Coriolis and frictional forces come into action. Moist air from the sea crosses the coastline and becomes more and more unstable as it is heated from below. Great vertical cloud development occurs, often accentuated by the presence of mountain ranges. Frequent heavy showers, thunderstorms and violent squalls take place over wide areas of the continent.

In winter, the situation is reversed. Air flows from the cooler land towards the sea. Subsidence is widespread and the air is drier.

Seasonal winds arising from temperature differences between the land and the ocean are called monsoon winds or simply monsoons. The name is derived from the Arabic word meaning 'season'.

19.6 **Monsoon regions**

Monsoon winds are most developed over southern and eastern Asia. In winter, the outflow from the Siberian anticyclone leads to a northwesterly wind across the Pacific coasts, and northerly or northeasterly flow across southern China, Burma and India. The air then moves across the Indian Ocean towards the equatorial trough lying south of the equator.

The Siberian anticyclone weakens in March, as the land temperatures begin to rise. However, the southwest monsoon of India does not arrive until about June. Over the western Pacific Ocean, the direction of the summer monsoon is southerly or southeasterly.

Monsoon winds also develop over other continents. In northern Australia, the 'dry season' occurs during the southern hemisphere winter, when the southeast trade winds extend over the region. When the trades begin to weaken, rising temperatures gradually lead to the establishment of a heat low.

Towards the end of the year the 'wet season' commences with isolated thunderstorms over the northern parts of the continent. When the northwest monsoon definitely arrives over the northwestern and northern regions of Australia in December, heavy rain may persist for several days. Thunderstorms, violent squalls and heavy precipitation usually reach a climax in January and February, when the equatorial trough is well established over the northern part of the continent.

Over Africa, the southeast trade wind crosses the equator and extends far inland as the southwest monsoon of the northern hemisphere summer. On the tropical east coast, the southeasterly or southerly winds of winter give way to northeasterly winds of the southern hemisphere summer. This effect is somewhat influenced by pressure changes over Asia.

Over North America, the monsoonal change is not so complete as over Asia. However, there is a marked change in the prevailing winds, particularly to the north of the Gulf of Mexico.

19.7 **Weather in the tropics**

The temperature in the tropics varies much less from summer to winter than it does in middle latitudes. This is mainly due to the fact that the midday sun is never far from the zenith.

If you refer to a map of the world, you will also notice that oceans cover most of the tropics. The temperature variations of the oceans are relatively small, being less than 3°C almost everywhere in the tropics. Even over the continents the annual range of the mean monthly temperature is less than 10°C. By contrast, it is 15°C or more in regions outside the tropics.

If you study the apparent motion of the sun, you will notice that most places in the tropics have the sun directly overhead twice during the year. This often leads to two periods of above-average temperatures. Sometimes, however, cloud and precipitation reduce the temperature for a certain period and produce a similar effect.

In general, very weak horizontal temperature gradients occur between the subtropical high pressure belts. As a result, it is impossible to detect well-defined fronts, such as occur in middle latitudes. Fronts entering the fringes of the tropics from higher latitudes dissipate rapidly, as the temperature contrasts between the air masses decrease.

The rainfall at any station depends on both synoptic scale and mesoscale effects. On the synoptic scale, depressions and troughs exert an important influence on the rainfall. However, a great deal of precipitation occurs in showers and thunderstorms arising from local effects, such as orography and surface heating.

19.8 **Convergence zones**

Although frontal activity is rarely evident, due to the weak temperature gradients in the tropics, widespread cloud development and rain frequently occur. Low level convergence in a depression or trough causes the air to rise, leading to cloud development and precipitation. This effect is accentuated if divergence takes place in the upper troposphere.

Convergence may also occur when two air streams approach each other from different directions. If moist tropical air is forced to rise, organised lines of cumulus or cumulonimbus clouds and belts of cirrostratus may develop in the vicinity of the convergence zone.

19.9 **Weather in the equatorial trough**

The equatorial trough is the region originally known as the doldrums. Although this is generally a region of light and variable winds, local thunderstorms may develop due to surface heating and orographic effects. Local rain squalls may therefore occur in this region.

Convergence zones within the equatorial trough may lead to more widespread cloud and precipitation. The tops of cumulus and cumulonimbus clouds spread out at higher levels and form sheets of altostratus and cirrostratus cloud. While thunderstorms occur in the convection cells, rain from the altostratus cloud may fall over wide areas. The width of the disturbed weather varies according to the scale of the convergence.

19.10 **The intertropical convergence zone**

Convergence occurs on a grand scale if the trade wind systems of each level meet in a narrow zone. This is known as the intertropical convergence zone (ITCZ). In general, the two systems are separated by a wide doldrum area, but in some regions the northeast and southeast trade winds approach each other more closely.

The intertropical convergence zone produces extremely bad weather conditions over a wide area. Vertical cloud development extends through the troposphere to the high tropical tropopause at altitudes of seventeen or more kilometres. The cloud base may be down to a few hundred metres, at times lowering to the surface. The zone of bad weather may sometimes be several hundred kilometres in width with general heavy rain, frequent thunderstorms and violent wind squalls.

19.11 **Diurnal and local effects**

Diurnal and local effects are very important in the tropics. They often exert a greater influence on the weather than some synoptic scale disturbances. Diurnal variations in temperature, wind and rain are often accentuated by orographical effects.

The variations in temperature which occur throughout the day depend largely on the prevailing wind direction. Where steady on-shore winds occur, the daily range of temperature is small. By contrast, the climate may be somewhat of a continental type, if the prevailing wind is off-shore. A relatively large range of temperature may then occur.

The sea breeze has an important effect on the temperature and weather at many tropical stations. It may also reinforce a prevailing on-shore wind. If the coastline is hilly, the sea breeze may be combined with the anabatic effect discussed in Section 14.4.

In regions where the air is moist and unstable, the sea breeze may lead to increased vertical motion, resulting in afternoon showers or thunderstorms. In a similar way, a land breeze may produce off-shore thunderstorms about dawn.

19.12 **Tropical disturbances**

Daily synoptic streamline charts show many features which are absent from mean annual or monthly charts. This is due to the smoothing, which occurs when average values are taken.

The synoptic systems found on daily charts include various types of eddies or vortices, wave patterns and convergence zones. These may be associated with convergence and divergence. Ascent or descent of the air takes place, resulting in many different types of weather on a synoptic scale. These, in turn, are modified by local effects in the various localities.

The term tropical disturbance refers to any feature of the flow that disturbs the basic tropical air currents. If tropical disturbances were absent, the flow could be represented by more or less smooth straight streamlines with steady speed.

The most vigorous tropical disturbances are the intense cyclonic storms that form over warm tropical waters. These are discussed in the next section.

19.13 **Tropical cyclones**

Intense cyclonic circulations on a synoptic scale occur in some tropical regions. These cyclonic storms develop much greater surface wind speeds than any other type of synoptic scale disturbance.

According to the locality in which they form, these large-scale cyclonic disturbances are known as hurricanes, typhoons or simply tropical cyclones.

Tropical cyclones develop in most tropical oceans at latitudes usually greater than 5° from the equator. They reach their greatest intensity while located over warm tropical waters. As soon as they move inland they begin to weaken, but often not before they have caused great destruction.

A tropical cyclone first appears on an MSL synoptic chart as a slight tropical disturbance. When a definite cyclonic circulation has been established and the wind has increased, but is still below gale force (34 kn), it is called a tropical depression.

If the wind force reaches gale force or higher, the disturbance is classed as a tropical cyclone. However, in some countries a distinction is made between a tropical storm (34 to 63 kn) and a hurricane, which is a severe tropical cyclone with even stronger winds.

The lowest MSL pressure of a tropical cyclone is frequently about 960 mb, but it is often much less. Often the depth is as great or greater than that of a deep extra-tropical depression. However, whereas the latter may measure as much as 3,000 km across, the diameter of a tropical cyclone is generally less than one quarter of that value, sometimes being smaller than 100 km.

Extremely steep pressure gradients are therefore possible and the surface winds may reach hurricane force. Torrential rain and thunderstorm activity with widespread cloud development accompany the storm.

A typically well-developed tropical cyclone consists of a more or less circular area of hurricane winds (i.e. over 63 kn), perhaps 80 km in radius. Outside this area the wind force decreases fairly rapidly. At a distance of 150 to 200 km from the centre the speed of the wind may be down to 30 to 40 kn.

The lowest pressure occurs at the centre, which is known as the cyclone eye. This region is sometimes only a few kilometres in diameter and is characterised by light winds, no rain and often very little cloud. It is sharply separated from the surrounding strong wind, rain and heavy cloud.

The cyclone eye is surrounded by a spectacular wall of cloud extending in a steep slope to the upper rim at an altitude of ten or more kilometres. Outside the wall cloud is the maximum speed ring, where the winds may exceed 300 km/h and towering cumulonimbus clouds produce torrential rain and thunder activity. These winds and associated rain are best developed in the forward quadrants of the cyclone.

The cloud and rain areas around the cyclone spiral inwards towards the eye wall. The rain areas can often be detected by radar and are known as spiral rain bands.

Tropical cyclones may move in any direction. However, very often they move westwards and slightly polewards after they first form.

The strong winds around tropical cyclones generate high waves. These travel outwards in all directions from the storm at an average speed of more than 1,500 km per day. This may be several times greater than the speed of the tropical cyclone itself. Hence, the arrival of heavy swell is a useful warning of the approach of a tropical cyclone, while the direction of the swell may provide an indication of the bearing of its centre.

A tropical cyclone derives its energy from the heat stored in the tropical waters and the latent heat released during the condensation of water vapour of the moist tropical atmosphere. It represents one of the physical processes in which large amounts of solar energy are eventually transformed into kinetic energy.

On the average, tropical cyclones last for about six days before they enter land or reach sub-tropical latitudes. Some can be detected for only a few hours, or perhaps a day or two; others are observed for as long as a fortnight.

The life history of a cyclone can be divided into four stages:

(a) The formative stage
(b) The immature stage
(c) The mature stage
(d) The decaying stage

However, the development may be terminated at any stage if the conditions are not favourable. Figure 19.4 shows the first three stages in the development of a tropical cyclone of typical dimensions. The diagrams refer to the southern hemisphere, but a mirror placed at the top of the diagram will indicate the northern hemisphere situation. In the latter case, the dangerous quadrant will then appear on the right hand front side, viewed from the centre along the direction of movement shown in diagram (c).

The four stages in the life of a tropical cyclone possess the following characteristics:

Formative stage
Tropical cyclones form in pre-existing disturbances. Deepening to a pressure below 1000 mb may take days or only a few hours. Winds reach gale force, usually in one quadrant only, and the eye forms.

Immature stage
This is reached when the central pressure falls below about 1000 mb. Winds reach hurricane force and spiral bands are formed. The area affected is small and the hurricane force winds may only blow within a 30 to 50 km radius. The surface pressure continues to fall. The cyclone may sometimes die within 24 hours of this stage.

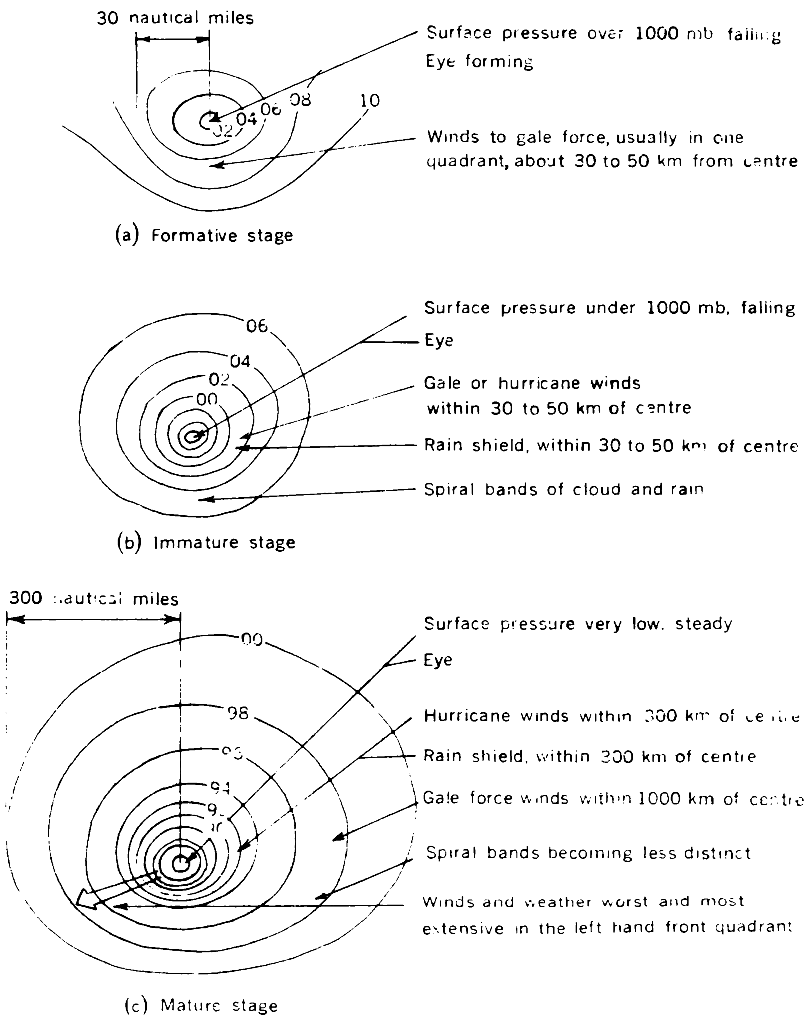
Mature stage
The surface pressure no longer falls but the area of the storm expands. Hurricane force winds may now extend as far as 300 km from the centre. In the southern hemisphere the gales and bad weather extend further to the left than to the right

of the direction of motion, with the worst weather occurring in the left forward quadrant. In the northern hemisphere the worst conditions occur in the right forward quadrant.

Decaying stage

At this stage the surface pressure rises and the area affected by the cyclone diminishes in size. In many cases it dissipates over land, possibly as a rain depression. More commonly, it moves to higher latitudes and becomes a sub-tropical cyclone, frequently of considerable intensity.

Fig. 19.4 Three stages in the life cycle of a tropical cyclone of typical dimensions (southern hemisphere)



In general, however, weather phenomena that occur in the tropics are either directly or indirectly related to events that take place in the atmosphere outside the tropics. In the next chapter we shall therefore investigate the atmosphere on a global scale.

CHAPTER 20

THE GENERAL CIRCULATION

In the preceding chapters we have been mainly interested in the state of the atmosphere at a particular instant and in a restricted region. However, it is necessary to study the atmosphere as a whole and over a long period of time, if we are to understand the various atmospheric phenomena that occur. This knowledge is essential if we are to predict the future state of the atmosphere.

One method of studying the atmosphere as a whole is to determine the average motion of the atmosphere over a period of many days. This motion is called the general circulation of the atmosphere.

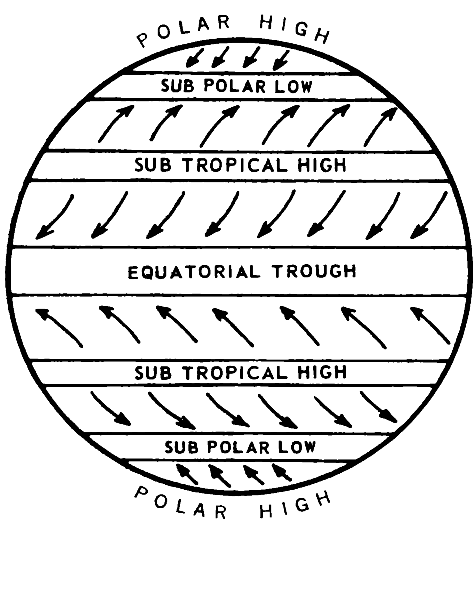
A time-averaged study of this type will necessarily smooth out many of the circulation patterns that appear on daily synoptic charts. However, it does form a useful starting point for studying many of the permanent and semi-permanent synoptic features that effect the weather in various parts of the world.

We shall first consider the essential features of the mean circulation of the troposphere and the lower stratosphere. Special features of the general circulation will then be discussed.

20.1 The mean circulation in the troposphere and lower stratosphere

Maps showing the average MSL pressure for January and July indicate that a region of low pressure exists at about latitude 60°, while a high pressure belt is located at

Figure 20.1 MSL circulation for a uniform surface



approximately latitude 30°. These features occur in both hemispheres and are known as the sub-polar low and sub-tropical high respectively.

There is also a region of relatively low pressure located in low latitudes. This is the equatorial trough.

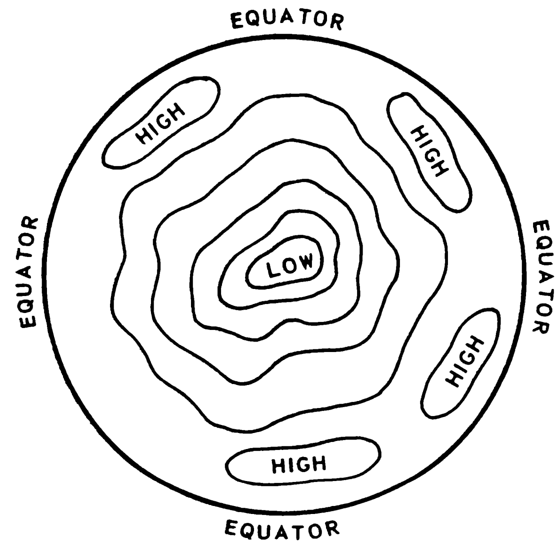
Figure 20.1 is the idealised mean pressure pattern which would apply if the earth's surface were uniform. In this case, there would be no distortions due to the uneven distribution of land and sea surfaces. The surface wind directions corresponding to this idealised situation are also shown.

A closer study of the sub-polar low pressure belt of each hemisphere indicates that it is composed of a number of cells. These low pressure centres are located somewhat more poleward in the summer of the hemisphere than they are in the winter.

Similarly, the sub-tropical high pressure belt of each hemisphere is characterised by a number of high pressure cells. Some of these occur in preferred positions and are called semi-permanent anticyclones. These also move polewards as the sun approaches during the summer of the particular hemisphere.

Upper air charts for the 700 mb constant-pressure surface indicate the synoptic patterns that occur at an altitude of about 3 km. Figure 20.2 shows mean 700 mb contours in the northern hemisphere summer. Note that the cellular patterns of MSL charts are no longer present in middle and high latitudes and that low contour values occur at or near the pole.

Figure 20.2 Typical mean 700 mb contours for one hemisphere



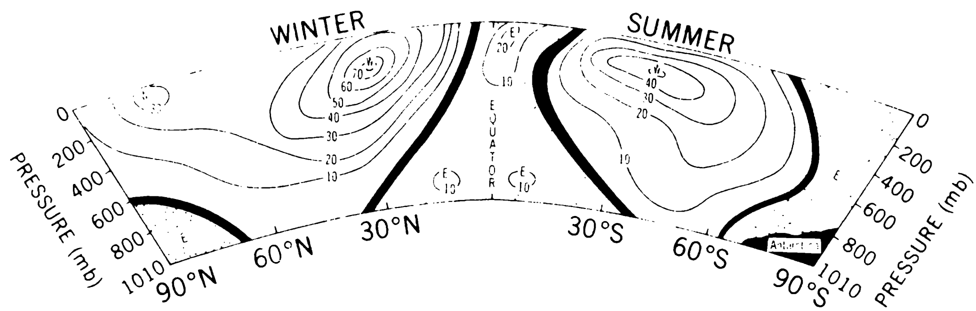
If the wind flows along the latitude circles, the motion is known as zonal flow. In practice, the wind direction varies from place to place, but it is possible to compute the average westerly or easterly component of the wind along any parallel of latitude. These are referred to as mean zonal winds.

It is possible to determine the mean zonal winds at various pressure levels in the troposphere and lower stratosphere. These are shown in Figure 20.3 as a zonal wind profile.

The centre of the sub-polar low is located near latitude 60° at the surface. It is tilted sharply polewards aloft and lies along the boundary between the westerlies and the polar easterlies.

The vertical axis of the sub-tropical high extends aloft from near latitude 30° at the surface. It leans equatorwards and is represented by the boundary between the westerlies and the tropical easterlies. In the lower levels, these easterlies are of course the mean zonal components of the trade winds.

Figure 20.3 The mean zonal flow (kn)



20.2 Jet streams

Figure 20.3 also indicates that the westerlies occupy a greatly expanded area in the upper troposphere as compared with sea level. The westerly component of the wind also increases in speed to about 200 mb (i.e. approximately 12 km) and then decreases aloft. The maximum wind speed of a winter mean cross-section is about double that of the summer.

A vertical wind speed distribution of this type indicates the presence of a jet stream. This is really a fast stream of air, confined to a relatively narrow 'tube'.

The World Meteorological Organization has defined a jet stream in the following way: 'The jet stream is a strong narrow current concentrated along a quasi-horizontal axis in the upper troposphere or in the stratosphere, characterized by strong vertical and lateral wind shears and featuring one or more velocity maxima. The speed of the wind must be greater than 60 knots'.

The jet stream indicated on the mean zonal profile is actually the result of the combined effect of two jet streams, which are observed on daily charts. One is the sub-tropical jet stream. This jet stream is located at about 200 mb in the vicinity of the sub-tropical high at about latitude 30°.

The other jet stream located on daily charts is the polar front jet stream. It appears just below the tropopause almost directly above the 500 mb position of the polar front.

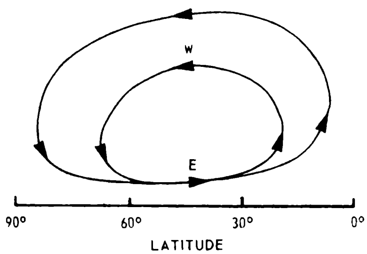
The latitudinal position of the polar front jet stream varies considerably on daily charts, as the polar front moves equatorwards and polewards. Periodically it merges with the sub-tropical jet stream, which is more persistent in its location. On mean zonal profiles, such as Figure 20.3 they also appear as a single merged jet stream.

20.3 Models of the general circulation

In order to study the general circulation it is useful to have a mental picture of its main features. This is known as a general circulation model.

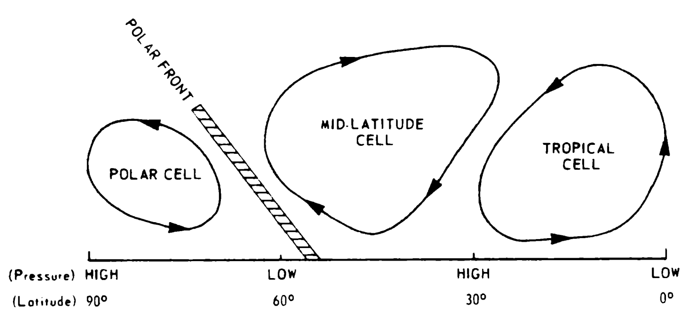
The first general circulation model was proposed by Hadley in 1735. It consisted of a single cell in each hemisphere similar to that shown in Figure 20.4. There are a number of scientific objections to this type of circulation and it certainly does not conform with the observational information available today.

Figure 20.4 Merdional cross-section showng Hadley's general circulation



In 1928 a three-cell circulation was suggested for each hemisphere by T. Bergeron, a Swedish meteorologist. This was later modified by C. G. Rossby in 1947. The mean features are shown in Figure 20.5.

Figure 20.5 Three-cell meridional circulation model



This model had the advantage that it provided for subsidence and relatively calm conditions over the mean position of the sub-tropical pressure belt. It also explained ascent and cloudy, rainy weather in the mean positions of the sub-polar low and the equatorial trough.

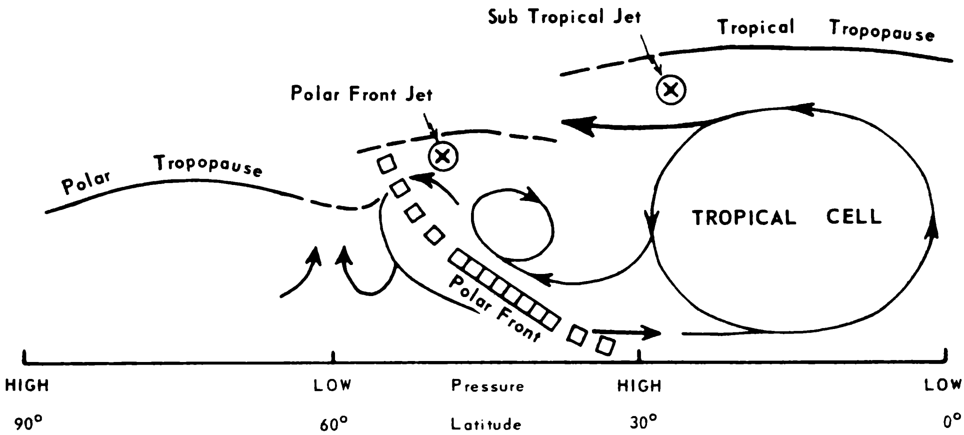
In the past two decades more upper air data has been obtained, and this has made it necessary for meteorologists to make some revisions to the three-cell model. For instance, in 1952, E. Palmén suggested that mean-circulation charts tend to smooth out some of the more important features of the circulation.

Figure 20.6 shows a revision of the winter meridional circulation suggested by Palmén. The polar front and sub-tropical jet streams are indicated by an encircled cross.

The three meridional cells or regions in Figure 20.6 may be described as follows:

(a) A tropical cell, similar to that of Figure 20.5
(b) An extratropical (or polar front) cell
(c) The regions poleward of the sub-polar low.

Figure 20.6 The Palmén model—meridional circulation in winter



The tropical cell conforms with our description of the tropical circulation in Chapter 19.

The extratropical cell is more complex than the tropical one. The migratory nature of the polar front complicates the picture. In addition, the mean cellular circulation is considerably weaker than in the tropical cell.

The broken portions of the polar front indicate elevations where the frontal contrasts are often weaker. This is partly due to meridional breakthrough currents which surge north and south across middle latitudes at these elevations.

In particular, mixing of the polar air occurs in the lower portion of the polar front, especially as it moves into sub-tropics. This process leads to the dissipation of polar fronts that move into sub-tropical regions.

Notice that a third breakthrough current occurs in the region between the polar and tropical tropopauses. Air from the troposphere enters the lower stratosphere in this way.

The meridional circulations poleward of latitude 60° are not fully understood. It appears from observations that the meridional circulations are weaker here than in middle and low latitudes. On the average, however, ascending currents can be expected to occur in the vicinity of the sub-polar low, in conformity with the low-level convergence in the vicinity of latitude 60°.

Some experimental approaches have also been made to the problem of determining the general circulation. 'Dishpan' models make use of a bowl containing a shallow layer of water. This is heated at the rim and cooled at the centre. It is then set rotating, while the circulation is observed by using drops of dye. Various circulations are noted according to the speed of rotation of the bowl, and some aspects are similar to those observed in the atmosphere.

More recently, use has been made of electronic computers to handle the enormous number of computations needed to determine the future state of the atmosphere from initial known conditions. It is still necessary to make some simplifying assumptions concerning the processes that occur in the atmosphere. However, realistic flow patterns have been achieved, and it is expected that in the future a much more detailed picture of the circulation can be obtained.

A great deal of the success with these studies will depend upon the availability of accurate data obtained from meteorological observers and from automatic weather stations located throughout the world. The Word Meteorological Organization is taking active steps to achieve this aim by means of the World Weather Watch.

List of abbreviations

μ	micron
°C	degree Celsius
CCL	convection condensation level
DALR	dry adiadatic lapse rate
ELR	environment lapse rate
ICAO	International Civil Aviation Organisation
ITCZ	intertropical convergence zone
K	Kelvin
kg/m^3	kilogram per cubic metre
km	kilometre
km/h	kilometre per hour
kn	knot
LCL	lifting condensation level
m	metre
mb	millibar
MCL	mixing condensation level
mm	millimetre
m/s	metre per second
MSL	mean sea level
p	pressure
SALR	saturated adiabatic lapse rate
T	temperature
WMO	World Meteorological Organization

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