MET. O.593

METEOROLOGY FOR MARINERS

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METEOROLOGICAL OFFICE

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With a Section on Oceanography

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FOREWORD

THE aim of this book is to present the elementary theory of modern meteorology in a way that is suitable for mariners, and to explain its practical application aboard ship. The chapters on ocean currents and ice are included because of their bearing on safety at sea. The section on oceanography has been specially written by Dr. J. N. Carruthers, of the National Institute of Oceanography, and has been included in this book because of its close connection with meteorology. We are indebted to him for the care and trouble he took in preparing these chapters. They were written in 1946 when Dr. Carruthers was serving in the Hydrographic Department of the Admiralty, and he has brought them up to date.

In this revised edition most of the material is broadly the same as in the First Edition, but certain amendments have been made in various chapters to bring it up to date. In particular, information about facsimile maps and about weather routeing has been added to Chapter 13, and further information about the application of meteorology to the care of cargo (adapted from an article published in *The Marine Observer* in October 1957) has been added to Chapter 1. A more up-to-date example of the Atlantic bulletin and its accompanying map takes the place of the old one in Chapter 13. Advice is also given in Chapter 13 about estimating wind force from a weather map. Maps showing frequency of Beaufort force 7 winds and above have been added to Chapter 7.

It will be noted that little reference is made in this book to meteorological instruments or to the making of meteorological observations at sea. The companion book, the *Marine Observer's Handbook,* deals in more detail with these aspects of maritime meteorology, and also with such subjects as ocean waves, meteorological, astronomical and optical phenomena and ice nomenclature, besides including a comprehensive collection of cloud photographs. The *Marine Observer's Handbook* is loaned free to each merchant ship of the British Voluntary Observing Fleet. It is also on sale at Her Majesty's Stationery Office.*

Many of the diagrams contained in this book are based upon observations made by observers aboard British Voluntary Observing ships. It is thus in some measure a tribute to the work of many thousands of voluntary marine observers.

Most of the weather systems illustrated in this book are those for the northern hemisphere.

Weather conditions have not the same significance in the navigation of a power-driven vessel as they had in the days of the sailing ship when a fast and safe passage depended almost entirely on every advantage being taken of wind and currents, and close attention to any sign of a change in the weather. Nevertheless, weather still has an important bearing on the economy of a voyage and must still be carefully watched to conserve fuel, and to avoid damage to the ship, her crew, passengers and cargo on the approach of adverse conditions. The care of cargo, the use of radar and a proper interpretation of a weather

* Up-to-date detail about meteorological codes (*see* Chapter 13) is given in Met. O. 509, *Ships' Code and Decode Book,* and in the *Admiralty List of Radio Signals,* Volume III; these codes are liable to be amended from time to time to meet changing international requirements.

forecast all require an understanding of the physics of the atmosphere. This is fully explained in Part I.

A study of the contents of this book should assist the ship's officer to become efficient at his job and to obtain his certificates of competency; and, it is hoped, will also encourage him to seek a wider knowledge of the interesting subjects of meteorology and oceanography.

From a study of the Board of Trade syllabus for Masters and Mates, students for First Mate are advised to study at least:

Chapter 1 (omitting pages 13–15), Chapter 2, Chapter 3 (omitting pages 35–41), Chapters 4 and 5, Chapter 7 (omitting pages 89–92), Chapters 8, 11 and 14, Chapter 15 (omitting detailed notes but study the ocean current chart).

Students for Master should make sure they study, in addition to the above: Chapter 3 (omitting pages 37–41), Chapter 9 (omitting pages 108–109 and 118–123), Chapters 10 and 13, Chapter 15 (concentrating on the ocean current chart; not the necessarily detailed notes).

Students for Extra Master should study the whole book.

Marine Division,

Meteorological Office.

PART I. THE METEOROLOGICAL ELEMENTS

CHAPTER I

THE ATMOSPHERE, ITS CONSTITUTION AND PHYSICAL PROPERTIES

Composition of the Atmosphere

The atmosphere surrounds the whole surface of the earth, both land and sea; just as life in the depths of the ocean is subjected to a pressure due to the weight of the sea above, so all of us, whether we be on land or sea, are living at the bottom of an ocean of air and are subjected to a pressure exerted by the weight of air above us. At sea level this pressure is about 15 lb/sq in. If we ascend a mountain or go up in an aeroplane, the pressure we experience at the new level is reduced, since the weight of air below our new level no longer contributes to the pressure exerted on us. Our bodies are adapted to the sea-level pressure of the atmosphere so that we are not conscious of it, but when we make a rapid ascent as in an aeroplane the reduced pressure affects the eardrums and renders breathing difficult owing to the reduction in the quantity of oxygen available. An increase of elevation from sea level to 2,000 ft will decrease pressure by about 1 lb/sq in, but this rate of decrease becomes less at higher levels. Roughly half the mass of the atmosphere lies below the level of the summit of Mt. Blanc (15,800 ft) and about two-thirds below the summit of Mt. Everest (29,000 ft). Because the upward rate of decrease of density with height becomes smaller and smaller the atmosphere becomes more and more tenuous at great heights and has no definite upper limit. However, even at 80–100 miles above the earth the air is still sufficiently dense for meteors to become white-hot and so visible when entering the atmosphere during darkness. The meteor is raised to white heat by the very rapid compression of the air ahead of it. This sudden compression is called an adiabatic compression: that is, it takes place without loss or gain of heat from outside sources. It is more fully explained in a later paragraph of this chapter.

The chemical composition of dry air is remarkably constant everywhere over the earth's surface and up to a height of at least 12 miles. Chemical analysis shows that the amount of each gas expressed as a percentage by volume of the total is as follows:

Gas	*Volume %*
NITROGEN	78·09
OXYGEN	20·95
ARGON	0·93
CARBON DIOXIDE	0·03

There are also very small amounts of NEON, HELIUM, KRYPTON, XENON, OZONE, RADON and perhaps HYDROGEN present.

While most of these gases play a vital part in the maintenance of all forms of plant and animal life on the earth, the exact composition of dry air is unimportant to the meteorologist, who is chiefly interested in the amount of water vapour that is mixed with it. The amount of water vapour in the air is limited solely by the temperature of the air; for every temperature there is an upper limit to the amount of water vapour which the air can hold. This amount increases from low to high temperatures. At temperatures below freezing the weight of water vapour which can be held is very small, while at high temperatures it can amount to 4% by weight of the air and water vapour mixture. Water vapour is constantly being added to the atmosphere by evaporation from the earth's surface, particularly from oceans, lakes and rivers, and is constantly being removed from the atmosphere by condensation resulting in precipitation in various forms, the chief of which are rain and snow.

Vertical Structure of the Atmosphere

Fig. 1.1 illustrates the vertical structure of the atmosphere but is drawn to scale only up to a height of 25 miles. If a uniform scale were used throughout, the important lower levels would be too overcrowded for the necessary detail to be shown clearly. The lowest region (shown by the close hatching) is called the TROPOSPHERE. Besides containing the greater mass of air and almost all the water vapour, the most important feature of this region is the decrease of temperature with height which persists through nearly its whole thickness. The average rate of decrease is about 1°F per 300 ft but considerable variations from this value are common. Sometimes shallow layers are found through which temperature increases with height. Apart from a few exceptions, which are dealt with later, the ordinary features of weather, including most forms of clouds and storms of all kinds, develop and expend their energies within the troposphere, the thickness of which can vary between 30,000 and 50,000 ft. At a level which is taken to be the upper limit of the region this rapid fall of temperature stops quite abruptly. For a considerable height above this level the rate of change of temperature with height is small, and the region is described as the STRATOSPHERE. The layer of transition, which may often be quite a sharp boundary between the troposphere and the stratosphere, is known as the TROPOPAUSE. The height of the tropopause varies from about 5 miles at the poles to 10 miles at the equator, and a curious result of this variation in height is that the stratosphere is colder above the tropics than at the poles. Over the equator the temperature at the tropopause is about −115°F; in temperate latitudes and polar regions it is on the average about −65°F. An increasing amount of flying is now being done in the lower stratosphere, partly because jet engines work more efficiently in the thin air at these levels, and also because this region is largely free from the hazards of convection clouds and the icing and turbulence associated with them. (*See* page 11, Table 1.1.)

Higher in the stratosphere a small quantity of ozone is found, mainly in a layer between 12 and 25 miles above the earth's surface. This ozone layer is important because ozone strongly absorbs radiation of certain wavelengths emitted by the sun, chiefly those in the ultra-violet region of the spectrum. If this ozone layer was not present an excessive amount of ultra-violet radiation would reach the earth with harmful effects upon many forms of life.

Direct measurements of temperature, humidity and winds can at present be made only up to about 20 miles; for higher levels we are dependent upon measurements by rockets and by inference from the way in which the sound of explosions is heard at great distances after reflection from a layer high in the stratosphere. These results can only be explained on the assumption that there

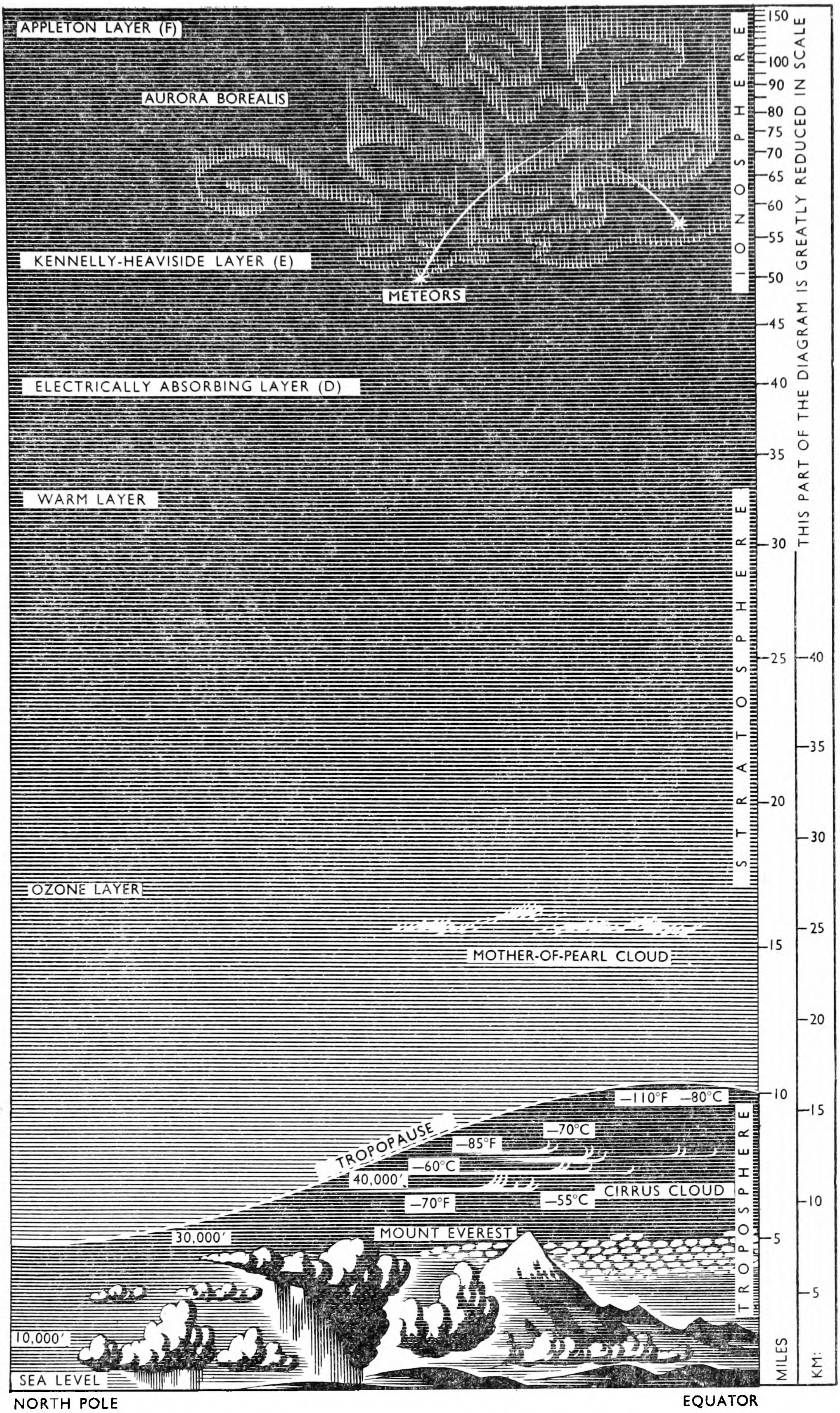


Fig. 1.1. The vertical structure of the atmosphere

is another warm layer with a temperature of about 100°F at 30 miles, a cold layer with a temperature about −100°F at 50 miles and that the temperature rises steadily in the region above 60 miles. The electrically conducting region above a height of some 50 miles is known as the IONOSPHERE where aurorae are located. In recent years much attention has been paid to the effects on radio and allied phenomena of these ionized layers, which have been named after various pioneer workers in this field.

Factors controlling Atmospheric Temperature

The temperature of the air at a given place and time depends upon a number of factors, the first of which is the amount of heat entering and leaving the atmosphere as a whole. Neglecting the minute contribution due to the leakage through the earth's crust of heat from the molten interior, all the heat reaching the earth's surface arrives in the form of radiation from the sun, and similarly all the loss of heat is in the form of radiation to space. A brief consideration of the properties of radiation is essential for an understanding of the way in which the atmosphere as a whole is heated and cooled. All bodies emit energy in the form of electromagnetic waves of very small wavelength which travel through space with the velocity of light. The hotter the body the shorter are the wavelengths which it emits, similarly cooler bodies emit a band of wavelengths which are generally longer than those emitted by hotter bodies. Thus the sun emits short-wave radiation in the wavelength 0·2 to 4μ, whereas the earth and its atmosphere emit long-wave radiation in the wavelength interval 4 to 50μ. (Radiation wavelength is measured in a unit known as a micron denoted by the symbol μ where 1μ is equal to 10^{-4} cm.)

Energy in electromagnetic waves travels through space in straight lines. No energy is lost by the radiation during its passage through space, which is another way of saying that space is entirely transparent. The atmosphere on a clear day is also nearly transparent to incoming solar radiation, which is thus only slightly depleted by absorption and scattering in its passage through the atmosphere before it reaches the land and sea surfaces where it is partly absorbed and partly reflected. The solar radiation which is absorbed heats the surface of the ground and the surface layers of the sea, which emit long-wave radiation appropriate to their temperature. This long-wave radiation is partly lost to space and partly absorbed by the atmosphere. Thus the sun's heat in the form of short-wave radiation reaches the earth's surface and has a very small heating effect on the atmosphere while passing through it. An understanding of this point is important because, as a result, the heating of the atmosphere is largely brought about by simple conduction of heat from the heated earth's surface to a thin layer of air in direct contact with the surface, combined with some absorption by the atmosphere of long-wave radiation from the earth's surface. From this layer the heat is carried upwards in ascending currents and eddies even to the top of the troposphere by the processes of convection and turbulence which will be described later.

Water present in the atmosphere as vapour or as cloud modifies the above process in an important way, since, while it absorbs only a small fraction of the incoming short-wave solar radiation, it strongly absorbs the outgoing long-wave radiation.

On cloudy nights this outgoing radiation is absorbed by cloud sheets which also emit radiation in all directions. The portion emitted downwards represents a gain of radiation to the ground and the atmosphere below the clouds, and consequently the fall of air temperature between the clouds and the surface is less than the fall on a clear night. Even on a clear night some of the long-wave radiation is absorbed by water vapour in the atmosphere and similarly re-emitted in all directions. The downward travelling portion helps to reduce the fall of air temperature in the same way. In both these cases the greatest reduction in the fall of temperature occurs close to the ground. This absorption of radiation by water vapour explains why the fall of temperature on clear nights in deserts and continental interiors in winter is so much greater than in localities having a maritime climate (e.g. western Europe), where the amount of water vapour held in air in which no cloud is present is often quite large. On cloudy days the temperature does not rise so high as on clear days, mainly because the upper surfaces of the clouds reflect a considerable proportion of the sun's radiation which reaches them back to space, amounting to between 56 and 81% in the case of dense clouds.

Since the mean temperature of the atmosphere as a whole does not vary appreciably from year to year, it can be concluded that the total energy of solar radiation absorbed by the earth and its atmosphere is balanced by the total energy of radiation emitted to space from the atmosphere.

Fig. 1.2 shows diagrammatically, for a summer midday, some of the processes involved in the exchange of heat in the form of short-wave radiation between the sun and the earth and long-wave radiation between the earth and space. The width of the beams also gives an approximate indication of the relative amounts of energy involved in the processes of transmission, scattering, reflection and absorption of solar and terrestrial radiation under the various conditions shown. The beams showing long-wave radiation refer to the net amount lost to space, that is, to the difference between the long-wave radiation from the earth's surface and the back radiation from the atmosphere. Scattering of short-wave radiation by air molecules has been omitted. (*See* Chapter 26.) Scattering, as its name suggests, is the term used to describe an effect of throwing out of the energy of solar radiation in all directions, produced by the molecules of the various gases in the atmosphere and by other suspended particles which are also very small compared with the wavelength of the short-wave radiation. Absorption of solar energy produces a corresponding rise of temperature in the absorbing substance; this is equally true for a gas as well as for liquids and solids.

Temperature Variation over the Earth's Surface

The primary cause of temperature differences over the earth's surface is the variation in the sun's altitude. If this were the only cause we might expect the highest temperatures at the equator and the lowest at the poles. In fact the distribution of temperature everywhere over the earth is controlled by several other agents, such as the distribution of land and sea, the amount of cloud cover through the year and the temperature of ocean currents. Dynamical factors such as the rotation of the earth on its axis, which has an important effect on the pressure distribution and the resulting system of winds, also play their part; while the variations of solar elevation and length of daylight through the seasons are everywhere responsible for the seasonal changes of temperature. The result is that the highest temperatures are found not at the equator, but over land which is covered by the sub-tropical high-pressure belts in summer. The lowest temperatures occur in winter in the heart of the biggest land masses,

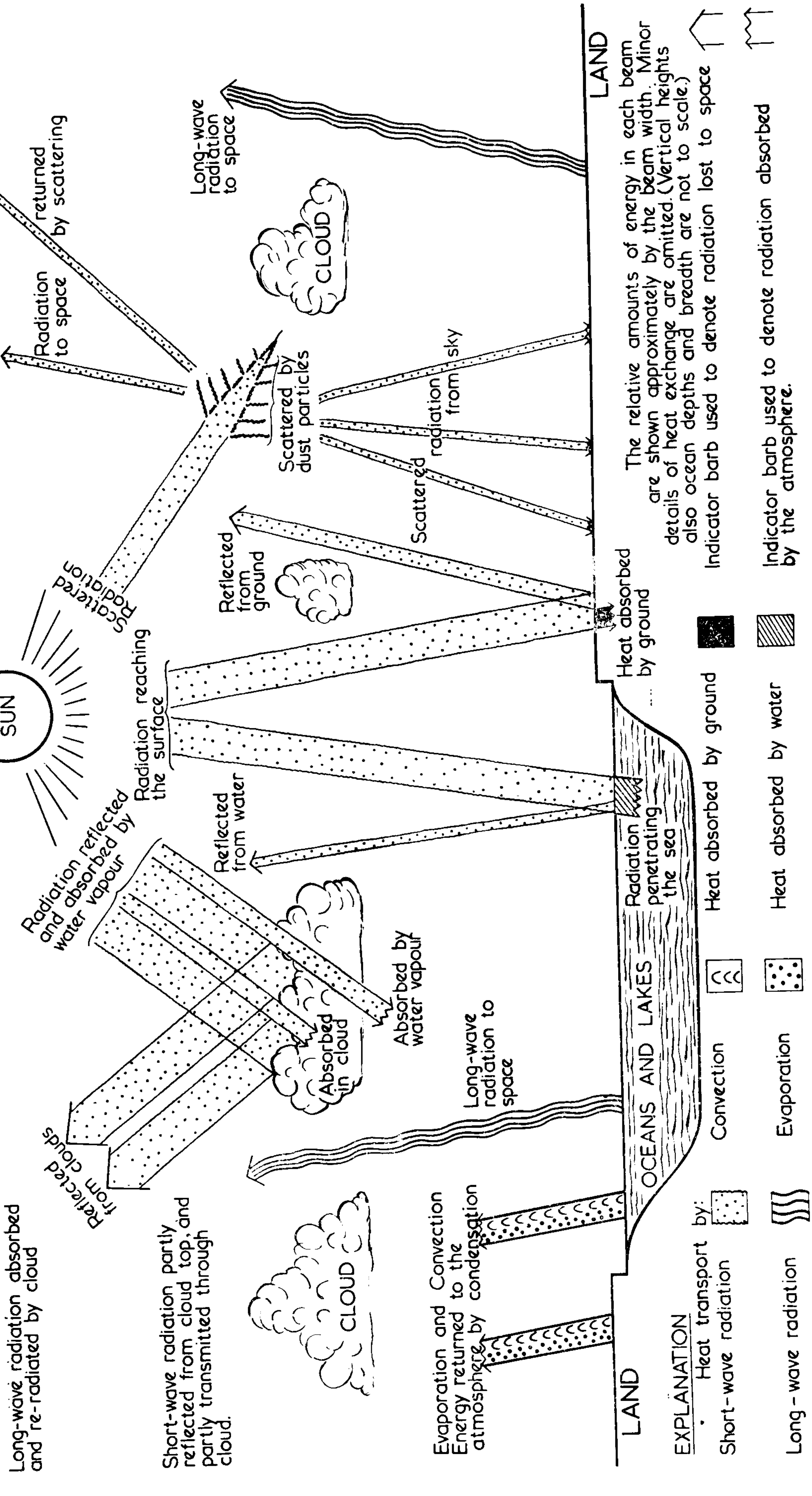


Fig. 1.2. Exchange of heat through the atmosphere and at the earth's surface

that is, in Siberia and north Canada. So far as is known temperatures in these regions are considerably lower than near the North Pole, though the small amount of data available suggests that the interior of Antarctica in winter is the coldest place on earth. (*See* page 94.)

Isotherms (Greek *Isos*, equal; *Thermos*, heat)

Isotherms are lines joining places having the same temperature. The distribution of mean sea temperatures, for January and July, over the surface of the oceans is shown by isotherms in Figs. 8.1 and 8.2, while the corresponding isotherms for mean air temperature are given in Figs. 8.3 and 8.4.

Water Vapour in the Atmosphere

Evaporation, or the escape of water vapour from a surface of water, snow or ice into the air above, goes on continuously all over the earth, so that the lower levels of the atmosphere always contain appreciable amounts of water vapour, especially over the warmer parts of the oceans. As stated earlier, the amount of water vapour which the air can hold is limited solely by the temperature of the air. At each temperature there is a definite maximum value of the amount of water vapour which the air can contain as vapour at that temperature.

The same facts can be described by saying that for every temperature there is a definite maximum value to the pressure which can be exerted by the water vapour, and that water vapour can only be retained as long as it does not exert a pressure in excess of this value. Any excess is condensed, i.e. returned to the liquid or solid form, usually first as cloud, but ultimately as precipitation, i.e. rain, snow, hail, etc.

Just as over a long period of years temperatures over the world as a whole appear to have altered little, so the evidence available suggests that there has been little, if any, sustained change in the annual amounts of cloud averaged over the whole earth through a long period of years. Therefore total evaporation in, say, a year is roughly equal to the total precipitation over the world in the same year.

The amount of water vapour in the air is important in many meteorological questions, besides directly affecting human comfort. One way of describing it would be as the weight of water vapour contained in a specified quantity of air. Fig. 1.3 shows how the maximum vapour pressure (or the vapour density expressed in grams of vapour per cubic metre) varies with temperature. It is important to note that the maximum amount of moisture that the air can hold increases more rapidly with temperature as the temperature increases. Between 20°F and 100°F the capacity of air to hold water vapour is roughly doubled with each 20°F rise in temperature. The lower layers of the atmosphere nearly always contain appreciable amounts of water vapour because evaporation, i.e. the escape of water vapour into the air from water, snow or ice surfaces, is going on for much of the time over most of the earth's surface. The amounts are greatest in air lying over or originating from the warmer parts of the oceans.

The saturation vapour pressures corresponding to ordinary temperatures are: at 32°F 6·1 mb, at 50°F 10·3 mb and at 68°F 23·3 mb, while at 212°F it is equal to normal atmospheric pressure, at which temperature boiling occurs (when atmospheric pressure is less than normal, boiling will occur at a lower temperature).

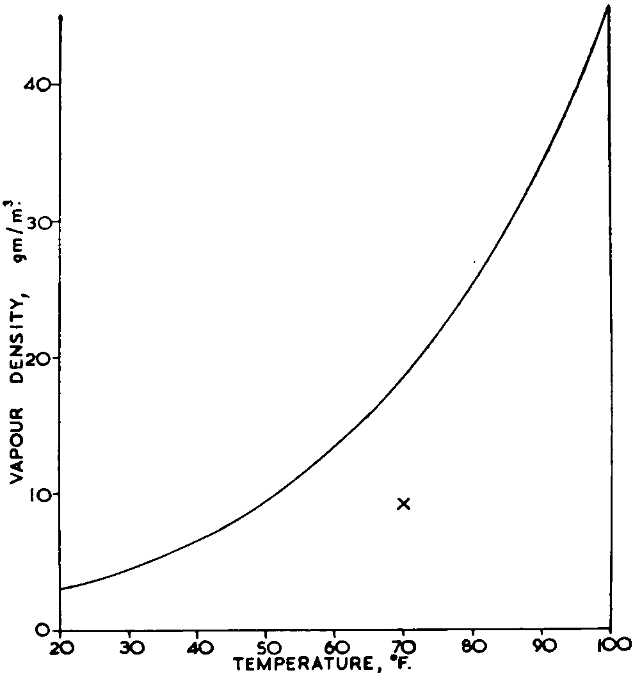


Fig. 1.3. Values of vapour density at saturation

Relative Humidity, Dew-point and Frost-point

A mass of air at a certain temperature can thus contain only a given maximum amount of water vapour. When this maximum figure is reached the mass of air is said to be saturated. More exactly, it is not the mass of air which is saturated, but the space occupied by the mass of air; this space is saturated when the water vapour contained in it is exerting the maximum pressure possible at that temperature without condensation taking place.

In general the air at a given locality contains rather less water vapour than is needed for saturation. When a meteorological observer finds that the air is holding half the quantity of water vapour needed to make it saturated at that temperature, he describes this fact by saying that the RELATIVE HUMIDITY of this air is 50%. In the same way, when his observation shows him that the air is saturated he can express the fact by saying that the relative humidity is 100%, i.e. the ratio (expressed as a percentage) of the actual amount of vapour held to that which the air could hold at saturation in this case is 100. Point X in Fig. 1.3 represents air at a temperature of 70°F and relative humidity 50%.

Apart from a feeling of dampness there is sometimes little to distinguish air of 100% relative humidity from air having a slightly lower relative humidity. However, this is only true if the temperature of this air is kept unchanged. If the air is saturated, a very small fall of temperature is sufficient to cause condensation; this will result in the formation of dew, if the air is cooled by coming into contact with a ground surface, at a lower temperature, of mist or fog if the air near the ground is being cooled, and of cloud if the air which is being cooled is at a higher level. The temperature at which condensation to water droplets occurs is called the DEW-POINT. (*See* the dew-point table on page 23.) If the dew-point is above freezing, condensation will be to water; if it is below freezing then moisture will be deposited on cold ground directly in the form of ice crystals (hoar frost). This process of direct deposition of water vapour as ice is known as SUBLIMATION.

The vapour pressure over a plane ice surface is less than that over a plane water surface at the same temperature; this is because the molecules are more tightly bound in the solid state. It follows that if the vapour pressure, e, is below 6·11 mb (the saturation vapour pressure at 32°F) it is possible to find two temperatures, one the dew-point, at which the saturation vapour pressure over *water* is equal to e, and the other, called the FROST-POINT, at which the saturation vapour pressure over *ice* is equal to e. When air is just saturated at freezing-point the dew-point and the frost-point have the same value, namely 32°F; but at lower temperatures the dew-point falls below the frost-point by an amount which increases as the temperature falls. For example, when the vapour pressure is 2·00 mb the dew-point is 6°F and the frost-point nearly 9°F. From the above it can be seen that there is a direct relationship between vapour pressure, vapour density, relative humidity and dew-point.

The formation of water droplets in the free atmosphere cannot take place in the absence of minute particles, known as nuclei. Such nuclei are numerous in the atmosphere though they vary greatly in numbers and origin. Salt particles, usually derived from sea spray, and minute particles of dust or sand, originating from deserts and from the products of urban areas, are the principal sources of nuclei. At levels in the atmosphere where the temperature is below freezing, condensation occurs upon such nuclei in the form of supercooled water droplets. Whether it is also possible for ice crystals to form in the free atmosphere, by direct sublimation of water vapour on suitable nuclei, or whether they always result from the freezing of liquid drops is still a controversial subject. It is a fact, however, that at temperatures down to about 14°F nearly all cloud particles are liquid droplets and that below about −40°F they usually consist of ice crystals. Between these two temperatures many clouds are mixed, i.e. they contain both liquid water and ice, but as the temperature falls ice crystals tend to predominate.

These facts are of great importance to the meteorologist concerned with cloud physics and the mechanism of precipitation. They may also be of significance to an airman because knowledge of the vertical distribution of temperature and of dew-point or frost-point is essential for forecasting cloud or (in war-time) the likelihood of condensation trails, while the composition of clouds has an important bearing on aircraft icing. Knowledge about frost-point and about supercooled water droplets in the upper air are admittedly of little practical importance to the mariner; reference is made to them here in order to 'complete the picture' of water vapour in the atmosphere.

Relative humidity varies widely in the lower levels of the atmosphere. In very dry air over desert regions values as low as 5 to 10% commonly occur, but over the oceans, where evaporation into the surface layers continuously occurs, such low values are uncommon and the air is often not far from saturation. At the ocean weather stations 'I' and 'J' in the north-east Atlantic* the annual mean value of relative humidity is 89%; as a comparison the figure is 79% for Kew Observatory in south-east England.

Hygrometers† (Greek *Hygros*, moisture)

Instruments for measuring humidity are called hygrometers. Relative humidity can be measured in several ways, but the meteorologist in practice makes use of only two methods. The simpler and more reliable method employs

* The station positions are 'I' 59° 00′N, 19° 00′W, 'J' 52° 30′N, 20° 00′W.

† *See* also *Marine Observer's Handbook*, Chapter 2.

a thermometer, the bulb of which is wrapped around by a piece of muslin which is kept moist by some simple device, usually by cotton threads which are tied round the muslin and run to a container holding distilled water. This thermometer is known as the WET-BULB THERMOMETER. Simultaneous readings of an ordinary thermometer (known as the DRY-BULB THERMOMETER), together with the wet-bulb thermometer, can be used to determine the relative humidity by means of tables. This is because there is a real physical relationship between the difference of the readings, dry bulb minus wet bulb, and relative humidity. In dry air the evaporation is large and results in a large cooling of the surface of the wet bulb, so that the difference between the dry- and wet-bulb readings is large. Correspondingly in damp air near saturation the difference between the dry- and wet-bulb readings is small, and decreases to nothing when the air is saturated and no evaporation is occurring at the wet bulb.

The other variety of these instruments in common use depends upon a property of human and animal hairs, by which they are able to absorb moisture and vary in length with the dryness or dampness of the atmosphere. The change in length of a bundle of hairs can be magnified through a system of levers and communicated to a pen and arm, which makes a record on a chart and thus enables a continuous record of the changes of relative humidity to be kept. In practice the readings of these instruments frequently have to be checked against those of dry- and wet-bulb thermometers, to guard against errors caused by changes in the properties of the hair and friction in the bearings.

Diurnal Variation of Temperatures

From sunrise until the sun is on the meridian, the rate at which any place on the earth's surface receives heat continuously increases with the elevation of the sun, in the absence of any change in the transparency of the atmosphere. Some increase is also due to the progressively shorter path of atmosphere traversed, though this effect is more variable, owing to changes in the atmosphere itself caused by cloud formation, drifting dust or other pollution. There is a corresponding decrease in radiation received from the sun from noon to sunset. All this time the earth is giving out heat by radiation, and this loss goes on increasing as long as the surface temperature is rising. The increase in temperature by day results from an excess of incoming solar radiation over outgoing terrestrial radiation. Under normal conditions the maximum temperature must be reached at some time after local noon, when a balance is achieved between the incoming and outgoing streams of radiation. In the British Isles, the maximum temperature usually occurs about 1400 (local time) in winter and 1500 in summer.

After sunset the earth loses heat by radiation less rapidly than by day because the temperature is lower. Both by day and night there is a net loss of energy in the form of long-wave radiation to space since the earth's surface radiates to the atmosphere rather more energy (about 50% more) than the atmosphere radiates back to the ground. This is true by day and night, even though the amount of long-wave radiation energy emitted by the earth is less by night because its temperature is lower. But at night-time there is no supply of incoming short-wave radiation to balance the net loss of long-wave radiation, hence the air temperature near the surface falls continuously until a minimum is reached when the incoming radiation equals the outgoing radiation just after sunrise*.

* See page 4 about the effect of cloud on outgoing radiation.

The short-wave radiation from the sun which strikes the surface of the land is absorbed in a very small depth of earth (*see* page 4), and consequently produces a large diurnal range of temperature, which is partly shared by the lowest layers of air near the surface. The sun's radiation which reaches the surface of the sea penetrates it to a considerable depth, with the result that the diurnal range of sea temperature and of the air in contact with it is very small even in the tropics.

Diurnal Variation of Relative Humidity

As we have already explained, if we increase the temperature of a mass of air in which the moisture content remains constant, the relative humidity of that air will decrease. Thus, even when the quantity of water vapour in the air at a locality does not change through the 24 hours, if there is a diurnal variation of air temperature there must be a diurnal variation of relative humidity, with the lowest relative humidity in this case coinciding with the time of maximum temperature. In general the amount of moisture in the air also varies through the day, being largest in the afternoon, so that the diurnal variation of relative humidity depends upon the daily variations of both temperature and moisture content, which are, however, both much less pronounced over the sea than over land.

Temperature Variation with Height

As the lower atmosphere, or troposphere, derives its heat indirectly from the sun, through heating of the earth's surface and of the layer of air in contact with it, it is not surprising that on the whole the temperature of the troposphere is highest at the bottom and decreases with height. The rate at which this decrease takes place is called the LAPSE RATE; on the average its value is 1 °F per 300 ft increase in altitude. However, the temperature of the troposphere does not always and everywhere decrease with height. The temperature at times, notably just over fog, increases with height. The term INVERSION is used to describe a layer through which temperature increases with height. Sometimes two or three such layers are found at different levels and an inversion is also present on occasions at the tropopause. The average temperature change in height is shown in Fig. 1.1.

Table 1.1. Pressures and Temperatures at various heights in the International Standard Atmosphere (ICAN)

Pressure mb	Height† ft	Temperature °F	Pressure mb	Height ft	Temperature °F
1,013·2	Surface	59·0	300	30,050	—48·2
1,000	436	57·4	200	38,630	—69·7
900	3,240	47·5	150	44,610	—69·7
800	6,390	36·2	100	53,040	—69·7
700	9,880	23·8	60	63,660	—69·7
600	13,790	9·8	40	72,090	—69·7
500	18,280	—6·2	10	100,000	—69·7
400	23,560	—25·0			

† The precise figures in the height column are due to the fact that they are converted to feet from the internationally used values which are in metres.

Adiabatic Changes (Greek *a-dia-baino*, not go through)

This is the name for the changes in temperature, pressure and volume which are produced in a substance when no heat is allowed to reach it or leave it while it is being compressed or expanded. Air, like other gases, is subject to certain natural laws. When air is compressed its temperature rises; conversely, when it is allowed to expand its temperature falls. The scientist Tyndall devised an experiment to show this, by placing a piece of dry tinder in a glass cylinder fitted with a piston. When air was rapidly compressed in this cylinder by quick pumping the tinder was set alight. No heat was given to the air from any external source; the temperature of the mass of compressed air was increased because the work done on it by the piston increased the energy of the molecules, making them move more quickly. The rise in temperature of a bicycle pump during vigorous use is a common illustration of the same effect.

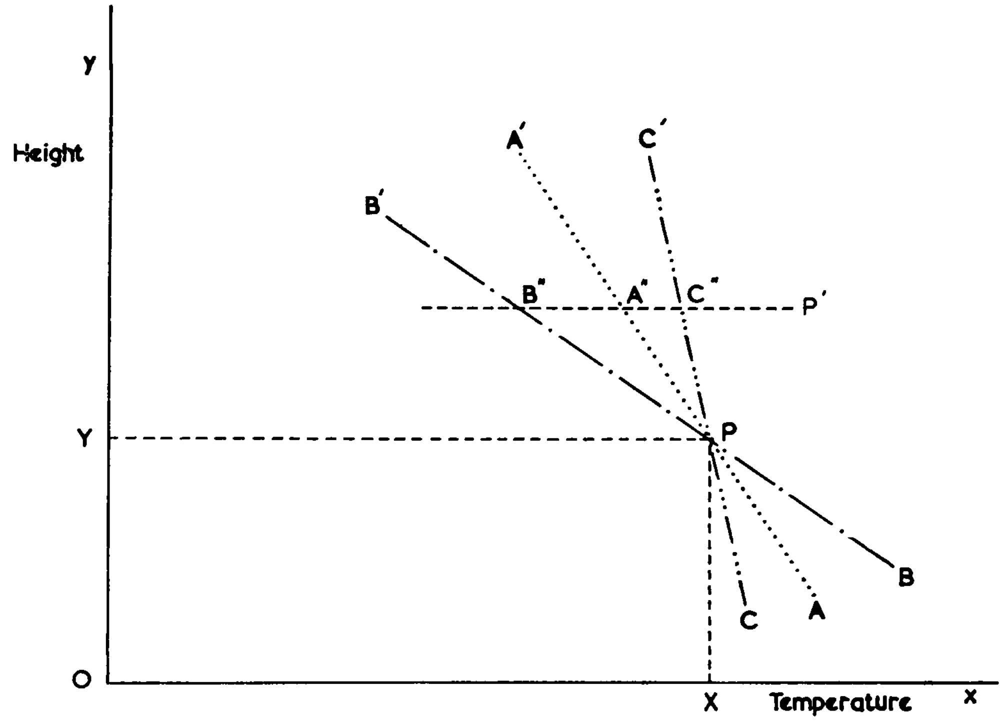
In one method of refrigeration a compressed gas is allowed to escape from a steel cylinder, and in the rapid expansion that follows it cools enough to cause freezing. In expanding against the surrounding pressure the gas uses energy derived from its own molecules, with the result that the average molecular velocity is reduced and the temperature of the gas falls. So long as these changes occur without any heat passing to or reaching the gas from an outside source, these changes are said to be adiabatic, and the increase or decrease of temperature thus produced is known as DYNAMICAL HEATING or DYNAMICAL COOLING.

We have already referred to the lapse rate of the atmosphere as the rate at which temperature decreases as we ascend through the atmosphere. In this case we are usually considering the value of the lapse rate which exists in the atmosphere over a particular place, and, provided the temperature and humidity at each level show negligible changes in a horizontal plane when short distances are considered, this lapse rate will vary little in the neighbourhood of the station and can be said to be representative of the air in the environment. However, air is a very bad conductor, so that in practice it is found that when a small 'parcel' of air near the surface is given a temperature slightly above that of the surrounding air at the surface, then it retains the additional heat which it has gained and this heat is not shared with the air in its neighbourhood. (Small parcels of air often gain more heat than the surrounding air; for instance, this happens on a sunny day wherever a patch of bare rock or soil an acre or two in extent is surrounded by grassland or woodland.) In this case such a heated parcel of air, being less dense than the adjacent air, will rise through levels where the pressure is progressively reduced so that it will undergo expansion. Since it does not lose (or gain) heat from the surrounding air while rising through it, the parcel of air is undergoing an expansion under adiabatic conditions. It is evident that the temperature of any parcel of air cooling under these conditions will fall with increasing height at the same rate, provided the parcel is dry, i.e. unsaturated. This rate of cooling is known as the DRY-ADIABATIC LAPSE RATE and has a value approximately 5·4°F per 1,000 ft of ascent.

When a parcel of saturated air similarly rises through its surroundings under adiabatic conditions, it cools at a rate which is rather less than the dry-adiabatic lapse rate. This is because, while the saturated air is rising, water vapour condenses from it, releasing latent heat which is absorbed by the rising parcel of air. Warm saturated air holds a large quantity of water vapour, cold air very little; in the warm air a large amount of latent heat is released, and thus the saturated-adiabatic lapse rate is not constant but varies between 2°F per 1,000 ft in the moist air of some tropical areas to almost 5·4°F per 1,000 ft at low temperatures and high levels in the atmosphere. On the average the saturated-adiabatic lapse rate at low levels in temperate regions is about one-half of the dry-adiabatic lapse rate.

Stability of the Atmosphere

The atmosphere is said to be stable when a parcel of air which is displaced slightly upwards or downwards is acted upon by a buoyancy force tending to restore it to its former level. Dry air is stable when its lapse rate is less than the value of the dry-adiabatic lapse rate; saturated air is stable when its lapse rate is less than the value of the saturated-adiabatic lapse rate. Similarly the atmosphere is said to be unstable when a parcel of air displaced slightly upwards or downwards is acted upon by a buoyancy force tending to move it up or down, i.e. further away from its former level. Dry air is unstable when its lapse rate exceeds the dry adiabatic, saturated air is unstable when its lapse rate exceeds the saturated adiabatic. When the lapse rate of dry air is exactly equal to the dry adiabatic through any thickness, and when the lapse rate of saturated air is similarly equal to the saturated adiabatic, there is consequently no restoring force acting upon a parcel of air when it is displaced either up or down, so the air in such a layer is said to be in neutral equilibrium.



.... APA′ (Dry-adiabatic lapse rate)

—·—· BPB′ (Actual lapse rate)

—...—...— CPC′ (Saturated-adiabatic lapse rate)

Fig. 1.4. Stability conditions of dry air

These facts are shown graphically in Fig. 1.4, where the variation of temperature with height (lapse rate) of a layer of air is shown by plotting height as ordinate **Y** and temperature as abscissa **X**. Suppose the actual lapse rate of a portion of the atmosphere is given by **BPB′** (in other words if the temperature at each height is plotted on the diagram, these points when joined form the line **BPB′**), while the dry-adiabatic lapse rate is shown by **APA′**. Consider the effect of displacing, under adiabatic conditions, an unsaturated parcel of air from its position at **P**. If the parcel is raised to the level **P′** its temperature will change at the dry-adiabatic lapse rate to become **A″**. But at this level the temperature of the surrounding air is **B″** which is lower than that of the parcel **A″**, which must therefore continue to rise because being warmer than its surroundings it is also less dense. Thus the air at **P** is unstable if displaced upwards. A similar argument will show that the air at **P** is unstable if displaced downwards and that the air at every level along **BPB′** is similarly unstable if displaced either upwards or downwards. The diagram can be also used to demonstrate the instability of saturated air for lapse rates exceeding the saturated adiabatic if we suppose **BPB′** denotes the actual lapse rate of a portion of the atmosphere which is saturated at every level (as often occurs in a fog bank or in thick cloud), and that **CPC′** represents the saturated-adiabatic lapse rate. Similarly, a line drawn to represent a lapse rate *less* than that denoted by **APA′** or **CPC′** would illustrate *stable* conditions.

By means of radiosonde observations (for description *see* the *Marine Observer's Handbook*, Chapter 3), supplemented at times by observations from special aircraft, a knowledge of the temperature and humidity of the air at various levels up to the tropopause, and frequently to a higher level, is obtained at fixed hours daily at a large number of stations in the British Isles, Europe, U.S.A. and many other parts of the world in temperate and some tropical regions. (*See* Fig. 1.5.)

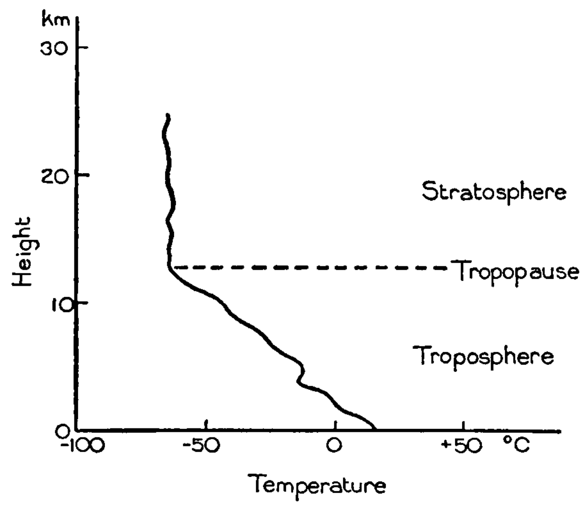


Fig. 1.5. Example of a temperature sounding through the lower atmosphere

An accurate knowledge of the temperature and humidity of the atmosphere at all levels from the surface to the upper troposphere on any occasion is of great value to the forecaster, because he is then able to make a definite statement on the type and amount of cloud which he expects will develop later on. For example, on a clear morning, if his information shows that the lapse rate is likely to exceed the dry adiabatic at low levels, he knows that volumes (i.e. 'parcels') of air will soon rise from the surface layers to a level where condensation occurs. From the argument in an earlier paragraph he knows that these parcels can continue to ascend freely through their surroundings at any higher level provided the lapse rate there is greater than the saturated-adiabatic lapse rate. So he will first estimate the time when the surface temperature will have risen to a value sufficient to establish a dry-adiabatic lapse rate up to a level at which a parcel of air, freely ascending from the surface, can become saturated. Next he looks at the temperatures at higher levels to see whether the lapse rate at these levels is greater or less than the saturated-adiabatic lapse rate. Normally it is much simpler to do this by plotting the temperature and humidity upon a TEPHIGRAM, which is a printed diagram upon which the curves of dry- and saturated-adiabatic lapse rates are inscribed for a number of dry-bulb temperatures at intervals of 5°F, in addition to co-ordinates based upon pressure, temperature and other meteorological elements which need not concern us here. From the curve which is obtained by plotting and joining these points upon a tephigram, the thickness of atmosphere through which the lapse rate exceeds the saturated adiabatic is readily determined. When the thickness is considerable, say several thousands of feet, any cloud developing in the layer will extend upwards to the top of the layer through which the lapse rate exceeds the saturated adiabatic. A lapse rate equal to or exceeding the saturated adiabatic must usually exist throughout a depth of about 10,000 ft to ensure that cumulonimbus clouds can become thick enough to allow showers to develop, similarly the saturated adiabatic layer must be about 15,000 ft deep to result in the formation of thunderstorms. If, as often happens, the lapse rate beneath the condensation level is less than the dry adiabatic, and the air there is not saturated, cumulus clouds will not form until a dry-adiabatic lapse rate has been established through the whole layer below the condensation level, which will not happen until this layer has received a certain amount of heating from below. This usually occurs after sunshine has had a chance to warm the surface layers for a few hours as described earlier, or when cold air is moving towards lower latitudes across a relatively warm sea surface. It is rare for the lapse rate in the free atmosphere to exceed the dry-adiabatic lapse rate except in a shallow layer; on the other hand the lapse rate frequently exceeds the saturated-adiabatic lapse rate through many thousands of feet in the lower atmosphere.

When the lapse rate is less than the saturated adiabatic, especially when there is an inversion of temperature a short distance above the ground, convection currents cannot occur and the formation of convection cloud is prevented. Inversions are common near the ground around sunrise following a clear night and, if the day following is clear, after several hours of sunshine the lowest layers of the atmosphere will often be warmed sufficiently to establish a dry-adiabatic lapse rate to a height above the condensation level. When this happens convection cloud will be formed as described earlier.

Subsidence

Stable and unstable air have been discussed above in terms of the behaviour of small parcels of air displaced upwards or downwards from their initial level. However, in some situations, as in the lower layers of a stationary anticyclone, there is considerable movement of air horizontally outwards over a large area. This air must be replaced by air slowly descending from higher levels: the process of descent is known as subsidence.

When air subsides, we can assume that it is warmed at the dry-adiabatic lapse rate. The moisture content is not changed by the descent, so the relative humidity must decrease as the temperature of the subsiding air rises, and subsidence areas are therefore occupied by warm, relatively dry air. Subsidence also alters the lapse rate, increasing the stability of originally stable air and resulting in the development of an inversion if the subsidence continues for long enough, e.g. for one or two days or longer.

Measurement of Air Temperature at Sea*

From the preceding discussion of radiation it can be understood that the accurate measurement of the air temperature a few feet above the sea surface is very difficult when the measurement has to be made on board the ship herself. The meteorologist requires the temperature of the air as though the ship were not there, but in practice the whole of the ship's structure above the water line is sending out long-wave radiation. Air reaching a Stevenson screen exposed on the ship's bridge is affected by this radiation, and may also be contaminated by being mixed with air from the ventilators serving the engine and boiler room. Therefore it is most important to hang the thermometer screen on the windward side, since it ensures that the heating effect of the ship, by radiation and conduction, on the air reaching the screen is reduced to the smallest amount possible. The effect of radiation can be further eliminated by the use of an aspirated psychrometer, provided the bulbs are protected by polished metal shields which reflect as much radiation as possible. When there is a following wind of speed equal to that of the ship, so that a relative calm exists aboard, the aspirated psychrometer is superior to the thermometer screen (provided the readings are taken as far out over the side of the ship as possible, and while the motor in the instrument is still running at full speed), since screen thermometers will be very liable to read wrongly owing to lack of ventilation.

Practical Value of Temperature and Humidity Information to the Mariner in Relation to the Care of Cargo

Knowledge about the physical processes involved in variations of temperature and humidity is of considerable value to the ship's officer in connection with care of cargo. Knowledge about sea temperature changes (*see* pages 94 and 95) is also valuable in this connection.

Appreciable damage, due to moisture effects, sometimes occurs to cargoes carried aboard ship. Some evidence of this is provided by contact with shipping interests and from enquiries received in the Meteorological Office for information about weather conditions at ports and along shipping routes on specified occasions when damage has occurred, but it is difficult to obtain any precise figures regarding the overall amount of damage. When a cargo is loaded into a ship its history prior to loading is important, although the ship's officer frequently has no means of knowing much about this. The temperature and weather conditions under which the cargo is loaded may 'come into the picture'.

Cargo damage includes such effects as mould formation, germination of grain, corrosion of metals, caking of sugar or chemicals, etc., and may arise either from 'hold sweat', i.e. water which condenses on cold deckheads, etc., and thence wets the cargo, or from 'cargo sweat', i.e. condensation directly on to the cargo itself. In studying the problem account must be taken of the special properties of cargoes, especially of those which are hygroscopic or which generate heat and release gases. Various sources and sinks of heat within the ship itself must also be considered, particularly when the cargo spaces are not

* *See* also *Marine Observer's Handbook*, Chapter 2.

insulated. These include the lower part of the ship's hull plates whose temperature is largely controlled by the sea temperature; the ship's structure above the water-line whose temperature varies due to radiation effects, from night to day and also from shaded to sunny sides; and hot or cold bulkheads adjacent to engine rooms or refrigerated spaces respectively.

Another important factor is the type of ventilation system provided. This may range from the old-fashioned cowl-type natural ventilation to the more modern forced ventilation with all-weather ventilator heads, or even, in a few cases, to full air-conditioning. Except where such expensive air-conditioning equipment is available, however, the ventilation of cargoes to remove unwanted heat or moisture can only be accomplished with the air that happens to surround the ship. If properly used, natural ventilation can give good results on the majority of occasions, but sometimes it is quite inadequate, e.g. when there is no wind relative to the ship. A number of rules of thumb have been developed, based on long experience, which have proved to be fairly sound when applied aboard such ships but they are far less reliable when applied on ships with more modern systems, for with these it is much easier to over-ventilate and to cause a lot of damage in a relatively short time. An officer who believed in ventilating whenever the weather was fine, for example, could do far less damage with natural ventilation than he could with mechanical ventilation. It is therefore essential that all officers should understand the relatively simple physical principles involved in correctly using any system of ventilation.

Principles of Ventilation

The main cause of damage to cargo arises from condensation of moisture from the air in the hold, either directly on to the cargo or on to the ship's structure, whence it may affect parts of the cargo. Thus the primary aims of ventilation are to remove excess moisture before it can do any damage and to avoid the introduction of such moisture in the ventilating stream. In some cases, of course, ventilation may be necessary to remove heat or gases generated by a particular type of cargo, and there are also circumstances in which all ventilation must be stopped because of heavy weather or other hazard. In general, however, the ventilating air must be capable of carrying away moisture, and a method is therefore needed to determine accurately whether or not it is doing so. This involves making measurements of the 'moisture content' of the air. (*See* page 10.)

Before dealing with the question of ventilation aboard ship it is desirable to refer to the special properties of different types of cargo, as these need to be clearly understood by ships' officers responsible for their care. Many cargoes of an organic nature such as wood, grain, wool, paper, cotton, etc., are hygroscopic, i.e. they can absorb water vapour from the air. For any particular hygroscopic substance having a specified moisture content and temperature, there is a value of relative humidity at which the surrounding air will be in equilibrium. If the actual relative humidity of the air is below this value, the substance will give up moisture to the air; if it is above this value then the substance will absorb moisture from the air. For a given temperature, this equilibrium relative humidity is higher the higher the moisture content of the substance; for a given moisture content it rises as the temperature rises. The moisture content of a hygroscopic commodity loaded in a warm climate is usually very large compared with the amount of moisture that can be taken up by the air in the hold of a ship, or even by a continuous stream of atmospheric air during the duration of a normal voyage, and it can therefore be assumed that it will remain more or less constant during such a voyage. Moisture equilibrium charts are available for some of the more common hygroscopic cargoes, and Fig. 1.6. indicates in outline the form which they usually take. (*See* also the dew-point table on page 23.)

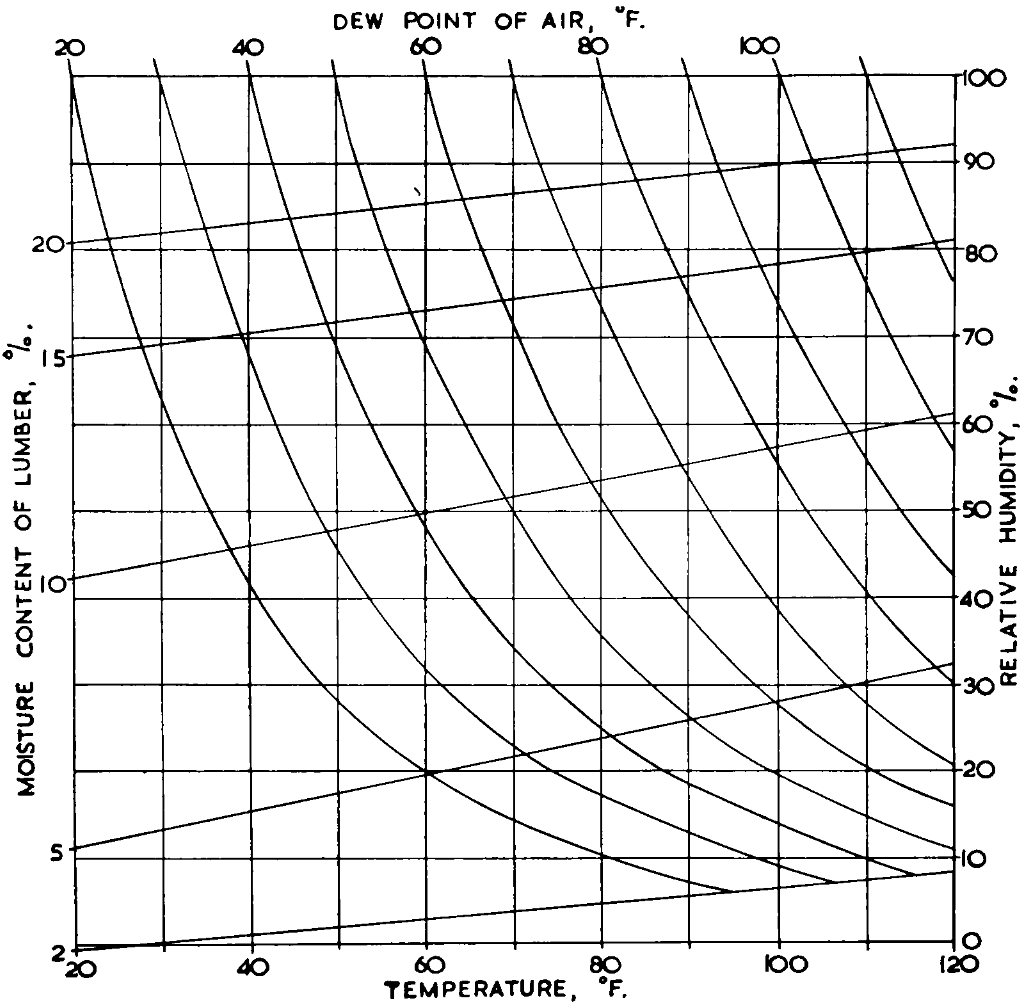


Fig. 1.6. Moisture equilibrium chart for lumber

It will be seen that this commodity (lumber), when it has a moisture content of 20 per cent by weight and a temperature of 80°F, will be in equilibrium at about 87 per cent relative humidity. This means that if the surrounding air is also at 80°F, it will take up or give up moisture according to whether its dew-point is below or above 76°F. If this lumber were in a confined space such as an unventilated hold, the dew-point of the air in the hold would fairly quickly become 76°F. In other words, in the absence of ventilation a hygroscopic cargo will control the dew-point of the air in the hold at a level which depends on the temperature and moisture content of the cargo. It follows, first, that it is desirable that hygroscopic commodities should be shipped with as low a moisture content as possible and, second, that such cargoes should be ventilated whenever possible to prevent the dew-point in the hold from building up so as to increase the probability of condensation on cool deckheads, bulkheads or plates. When dealing with hygroscopic cargoes it also follows that it is particularly important to consider all possible sources of heat and any cold surfaces in the hold. The presence of a hot bulkhead near a hygroscopic substance will raise its temperature and hence its equilibrium relative humidity, so causing the dew-point of the surrounding air to rise. This moisture may then be condensed on the colder parts of the cargo or on a cold bulkhead or deckhead. Insulation of such a cargo from any hot or cold surfaces is thus very desirable.

Non-hygroscopic cargoes neither absorb nor liberate moisture, and if they are stowed in a dry unventilated hold, the dew-point of the surrounding air will remain more or less constant even though its temperature may rise considerably during a voyage. Common cargoes of this type are steel products and canned goods. Steel products, because of their high heat capacity and good thermal conductivity, warm up relatively slowly as the temperature of their surroundings increases. Unless well protected by grease or some other coating they are therefore specially liable to condensation and rust damage. Such a cargo loaded in a cold climate will, as the ship sails into warmer waters, retain a temperature lower than it surroundings but if the hold is not ventilated the dew-point in the hold will be lower still and all will be well. If ventilation is begun, however, with outside air having a dew-point higher than the cargo temperature, then condensation on the metal surface will immediately result. If there is a hygroscopic cargo in the same hold there may also be trouble because the dew-point in the hold will then rise as the temperature increases and there is then some risk of condensation on the colder metal cargo, whether the ventilators are operated or not. Thus canned goods packed in boxes made of wood having a fairly high moisture content are specially prone to damage in this way. It has been known for steel products to be corroded due to moisture originating in the wooden dunnage used to protect them. In general, hygroscopic and non-hygroscopic cargoes should not be mixed in the same hold because they require different treatment in regard to ventilation. The same sort of thing would apply to two hygroscopic cargoes with appreciably different equilibrium relative humidities, arising from differing properties and moisture contents.

Returning to the ventilation problem aboard ship, it has already been stated that if ventilation is to be useful the ventilating airstream must carry away moisture or heat or both, and that to find out whether ventilation is necessary or is likely to be harmful, in any given situation, the temperature and dew-point of the airstream must be measured. First these quantities must be measured for the outside air, this being the air which will be drawn into the ship when the ventilating system is put into operation. Assuming that the ship's officer is able to measure the temperature and dew-point of the atmospheric air accurately from the weather side of the ship's bridge, he will have some valuable information to go on. For example, if the air is very warm and has a small dew-point depression it must contain a large amount of moisture and, unless it is considerably warmed, cannot be expected to take up much additional moisture; if cooled slightly it is likely to deposit moisture. On the other hand, if the air shows a large dew-point depression it is capable of taking up a considerable amount of water vapour and is likely to be suitable for ventilating unless the cargo is a cold one. Before a decision can be reached about ventilating, however, it is necessary to have a fairly good idea of the temperature of the cargo. Direct measurement of cargo temperature is difficult. Conditions in a hold might be measured by placing recording thermo-hygrographs either directly on the cargo or in spaces between the cargo, but this method would be of little value for day-to-day use because cargo spaces are often inaccessible and the readings taken by one or even two such instruments may not be representative of conditions in cooler and more critical parts of the hold. A somewhat better method, which would give readings of air temperature and humidity in any desired part of the hold, would be to use simple, distant-reading, electrical thermometers previously placed in the cargo and wired temporarily to junction boxes in the ventilator. A portable balance bridge indicator could then be taken round to each ventilator, say once a day, and plugged in by a ship's officer to get the necessary readings. Such an arrangement would be simple and inexpensive. Humidity readings throughout the voyage would be practicable by using a large enough water container. In general the temperature of the air in contact with the cargo will tend to be the same as that of the surface of the cargo and it will be noted that the methods just described would, strictly speaking, measure the condition of the air and not that of the cargo. It is believed that some ships' officers take hold temperatures simply by lowering an ordinary thermometer down a ventilator or perforated bilge sounding pipe. Such measurements are subject to considerable error but will serve as a rough guide in estimating the probable cargo temperature, particularly if taken in conjunction with information about the conditions under which the cargo was loaded, its nature and stowage, recent weather conditions, including sea temperatures (particularly if dealing with the lower hold) and whether proceeding towards warmer or colder conditions. When considering ventilation of holds, it should be borne in mind that the deck head under an unshielded deck will be very quickly affected by a change of temperature on deck; a sudden fall of air temperature may very quickly cause condensation to occur under the deck head in such a ship. If the deck is sheathed or otherwise shielded, then this effect will be appreciably less.

As a general rough guide, it is best not to ventilate if the *hold temperature* (or estimated cargo temperature) is *below* the *dew-point* of the outside air. If, however, the *hold temperature* (or estimated cargo temperature) is appreciably *higher* than the *dew-point* of the outside air, then it will generally be safer to ventilate.

If, however, the reading of the dew-point of the air in the hold can be obtained, more accurate guidance can be given, because moisture will only really be removed from the hold if the dew-point of the air in the hold is higher than the dew-point of the outside air. If it is not practicable to measure the dew-point in the hold in the case of a hygroscopic cargo, it should be possible to estimate it if the temperature and moisture content of the cargo is known and a moisture equilibrium chart for the commodity is available. Once ventilation has been started, however, the problem becomes much simpler because it is then possible to find out whether the ventilating stream is removing or adding moisture by direct measurement of its dew-point at or near the point of discharge. It is obvious that if the dew-point of the exhaust air is higher than that of the atmosphere then moisture is being removed, and vice versa. If a reasonably close watch can be kept both on the atmospheric dew-point and on that of the emergent airstream then the ventilation can be properly controlled. This is also the main conclusion reached in a booklet which was prepared on the subject of cargo ventilation by a working group of the Commission for Maritime Meteorology of the World Meteorological Organization and which was published in 1957. The procedure recommended for moisture control is therefore as follows:

(i) Measure the dew-point and temperature of the outside air.

(ii) Place the ventilation system into operation, unless the estimated cargo temperature is lower than the outside dew-point and the cargo is one likely to be seriously affected by a short period of condensation, e.g. metal goods.

(iii) Shortly afterwards measure the dew-point of the air flowing out from the ventilators to make sure that it shows a higher dew-point than that of the outside air. Unless it does there can be no removal of moisture from the hold and it will be useless or perhaps damaging to continue ventilation.

(iv) At reasonable intervals repeat the measurement of dew-point in the discharge stream to make certain that it is still higher than that of the input air.

(v) At longer intervals (6 hours) re-check the dew-point of the atmosphere around the ship to provide a new basis for comparison.

Some typical situations are considered below:

(*a*) Ship proceeding from a cold to a warm climate. Atmospheric temperature 70°F, dew-point 60°F. Temperature when cargo was loaded about 45°F.

If the cargo is a non-hygroscopic one, e.g. steel products, then it will still have a relatively low temperature, and the dew-point in the hold, assuming it to be dry, will also be quite low, certainly not more than 45°F. Ventilation will thus increase the moisture in the hold and, if the cargo temperature is still below 60°F, serious condensation, resulting in corrosion, may occur. On the other hand, ventilation will accelerate the rate of warming of the cargo and this will reduce the risk of condensation when unloading at the port of discharge: thus if a period occurs when the cargo temperature has risen sufficiently to exceed the dew-point of the outside air, then ventilation will be advantageous. (Re-circulation of the air within the hold, if this is possible, will serve the same purpose.)

If the cargo is hygroscopic it will also warm up slowly but in this case the hold dew-point will rise as well. Ventilation before the dew-point in the hold exceeds the outside dew-point will serve no useful purpose and might cause some condensation on the cold cargo.

(*b*) Ship proceeding from a warm to a cold climate. Atmospheric temperature 75°F, dew-point 60°F. Temperature when cargo was loaded about 85°F.

If the cargo is non-hygroscopic then the risk of condensation on it is small because it is warm and will cool down only slowly. There is some risk of condensation on the ship's sides and deckhead as these cool but this will usually be prevented by light ventilation. Ventilation should therefore be started and continued unless at any time the discharged air shows a lower dew-point than the atmospheric air.

If the cargo is hygroscopic then the air in the hold will receive large quantities of moisture and perhaps some heat from it, especially if its moisture content is high, or if it was loaded in a wet condition. Thus there is danger of heavy condensation on the ship's hull and deckheads. Vigorous ventilation is necessary so long as the dew-point of the discharge stream remains higher than that of the atmospheric air.

It will have been noted that hygroscopic cargoes being shipped from a warm to a cold climate should always be ventilated by air of lower dew-point to prevent the possibility of hold sweat, but that in practice this will not always be possible either because the atmospheric dew-point is too high or because of bad weather. This is where a ship equipped with air-conditioning equipment has an advantage (which needs to be considerable in view of the installation cost). But nothing will be effective, not even the most expensive apparatus, unless the ships' officers understand the principles involved.

In addition to the condensation risks referred to above, there is always a risk of damage to steel goods by salt particles in the air admitted to a ship's hold through the ventilators; this seems to indicate that it may be unwise to ventilate such products unless circumstances are particularly favourable to do so, or unless the steel goods are carried in the same compartment as hygroscopic goods.

Making the Necessary Measurements of Temperature and Dew-point on Board Ship

The difficulty of making the necessary measurements of temperature and dew-point with the required accuracy has already been mentioned. The average ship has no wet-bulb thermometer, and even if both dry-bulb and wet-bulb thermometers are provided they may be of doubtful accuracy and may perhaps not be correctly exposed or shielded from radiation effects or have a wet bulb that is properly cared for. Thus reliable measurements are often unobtainable. It is essential to proper ventilation management that ships' officers should have the means of making the required humidity measurements easily and accurately. Selected Ships are supplied on loan by the Meteorological Office with an accurate hygrometer (wet- and dry-bulb thermometer) in a portable screen. Assuming that the instructions about the care of the thermometers and exposure of the screen are complied with, accurate observations are thereby obtainable. This instrument is, however, intended for measuring the temperature and humidity of the outside air for meteorological purposes and it would be difficult to get readings of the 'discharge stream' of air from ships' ventilators by this means. A useful instrument is a portable aspirated hygrometer of the Assman type—driven by clockwork, mechanically or electrically; being readily portable it can easily be taken to the weather side of the bridge to get the atmospheric readings or to the discharge stream of accessible ventilators—but it needs a lot of patience in order to achieve accuracy, i.e. it may take two to three minutes before the reading becomes 'steady'. The possible use of electric thermometers in the ship's holds has already been mentioned; these could also be used for getting atmospheric readings on deck (on the weather side of the ship). Tables are available which give the various humidity parameters, relative humidity, dew-point, vapour density, etc., in terms of the dry-bulb temperature and the depression of the wet bulb. The tables differ somewhat according to whether they are to be used with thermometers in screens, as used at British meteorological stations and on British selected ships, or with aspirated psychrometers. In the former an average air speed past the bulbs of about 2–4 kt is assumed while in the latter air is drawn past the bulbs by means of a fan at a speed exceeding 7 kt, the critical speed above which the wet-bulb reading is independent of the rate of ventilation. The aspirated psychrometer is capable of the more accurate results but is generally more expensive. A dew-point table for use with an aspirated psychrometer is given below. It might be thought that the directly measured wet-bulb temperature and wet-bulb depression could be used in ventilation problems instead of the dew-point and dew-point depression, which have to be derived from tables. The wet-bulb temperature, however, is not by itself an accurate index of the amount of water vapour in the air, thus the wet-bulb temperature will rise simply as a result of an increase in temperature, without any addition of moisture. Moreover, the table shows that for the same wet-bulb depression colder air shows a greater depression of dew-point, so that the wet-bulb depression is not a good indicator of the margin for cooling before there is any danger of moisture being deposited.

Table 1.2. Dew-Point Temperature

Air temperature °F	Depression of wet bulb														
	1	2	3	4	5	6	7	8	9	10	11	12	13	14	15
20	15	10	5	—2											
25	21	17	13	8	2										
30	27	24	20	16	11	6	—1								
35	33	30	27	24	20	16	11	5	—4						
40	38	35	33	30	27	24	20	16	11	4	—4				
45	43	41	39	36	34	31	28	24	21	16	11	5	—4		
50	48	46	44	42	40	37	35	32	29	25	21	17	12	5	—3
55	53	51	50	48	46	43	41	39	36	33	30	27	23	18	13
60	58	57	55	53	51	49	47	45	43	40	38	35	32	28	25
65	63	62	60	59	57	55	53	51	49	47	45	42	40	37	34
70	69	67	66	64	62	61	59	57	55	53	51	49	47	45	42
75	74	72	71	69	68	66	65	63	61	59	58	56	54	52	49
80	79	77	76	74	73	72	70	68	67	65	64	62	60	58	56
85	84	82	81	80	78	77	75	74	72	71	69	68	66	64	63
90	89	87	86	85	84	82	81	79	78	76	75	73	72	70	69

Use of Climatological Information

It has been mentioned already that hold sweat occurs when the temperature of the ship's hull or the deckheads falls below the dew-point of the air in the hold. It is therefore evident that the danger of this occurrence must be greatest in those areas where a large fall in air temperature or in sea temperature or a large rise in dew-point may occur over a relatively short distance. (*See* the air and sea temperature maps between pages 96 and 97.) Such areas can be regarded as danger zones for cargo damage and they can be defined by a study of the variation of these elements along the main shipping routes.

Unfortunately dew-point information has not yet been portrayed in these atlases at all because it is only since 1953 that dew-point has been punched on to the Hollerith cards that are used for marine climatological work and it takes many years to accumulate enough data for portrayal on a map. However, using the data at present available, mean dew-point isopleths in °F have been computed for each month for each of the oceans. These are being included in the Routeing Charts at present (1965) being published by the Hydrographic Department of the Navy. Some work on the subject has also been carried out in the Marine Division of the Meteorological Office and it is hoped to make available in due course a series of diagrams for each main shipping route showing the mean and extreme values of air temperature, dew-point temperature and sea temperature experienced along that route in each month of the year. Figs. 1.7 and 1.8 show the kind of way in which the data might be presented but the eventual product may well be different from this. The upper set of curves shows the variation of the mean values for January along the main Atlantic shipping route. The lower set of curves indicates possible extreme falls in air and sea temperature and rises in dew-point temperature in the next 24 hours for a ship steaming towards the west at 10 knots. The lower curves have been derived from the upper ones plus the corresponding 5-percentile maximum and minimum curves* (only short sections of the latter curves for sea temperature are shown in order not to confuse the diagram). In whatever way the data are presented the object would be to show clearly the areas in which the risks of rapid falls in sea or air temperature or of a rapid rise in dew-point temperature are greatest. It can be seen from Fig. 1.7, for example, that in January on this route a rapid rise in dew-point is most likely to occur in about longitudes 40–45° and the greatest risk of large falls in both sea and air temperature occurs towards the American coast.

Variation of mean air temperature, sea temperature and dew-point temperature along route United Kingdom to U.S.A., January

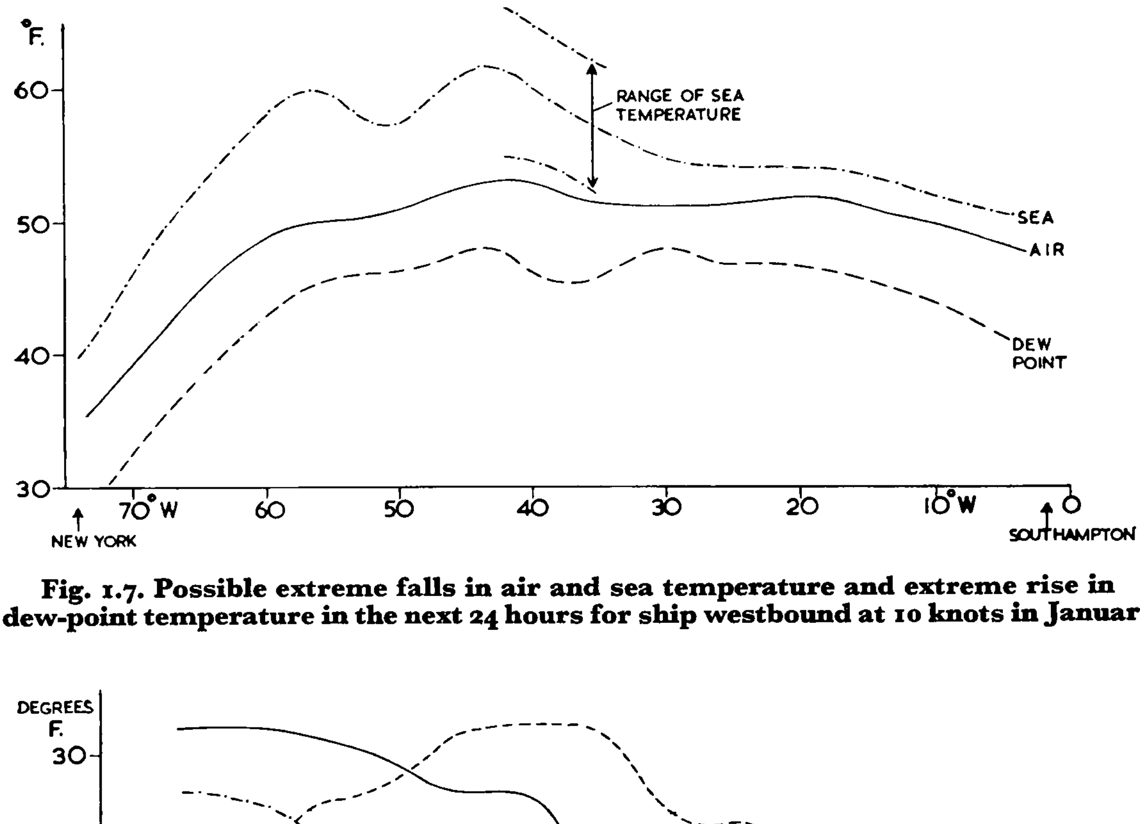


Fig. 1.7. Possible extreme falls in air and sea temperature and extreme rise in dew-point temperature in the next 24 hours for ship westbound at 10 knots in January

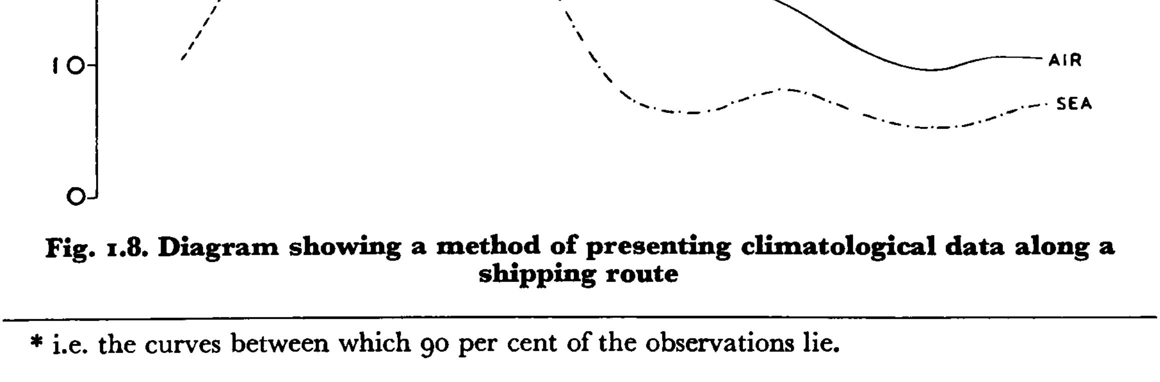


Fig. 1.8. Diagram showing a method of presenting climatological data along a shipping route

* i.e. the curves between which 90 per cent of the observations lie.

CHAPTER 2

PRESSURE

Units of Pressure. Millibars

The atmospheric pressure at sea level averages about 15 lb/sq in., which represents the weight of a vertical column of air of 1 sq in. cross-section extending to the top of the atmosphere. Since the air is fluid, and for practical purposes in statical equilibrium, it follows that the pressure at any point is the same in all directions. We may therefore regard the pressure as supporting the vertical air column. The atmospheric pressure can similarly be made to support a column of heavy fluid, as in the mercury barometer, where pressure is measured by noting the length of such a column. In the past, not only was atmospheric pressure measured in this way but the result was recorded as a length, such as 30 in. (approx. 760 mm) of mercury, and all barometers, mercurial and aneroid, were graduated in these units. One objection to recording pressure in this way is that it makes the recorded value dependent on the density of the fluid used, in this case mercury. In addition, there is risk of confusion of units; pressure is not normally expressed as a length but as 'force per unit area'. To avoid these difficulties a unit of pressure has been introduced. This unit, known as the MILLIBAR, is defined as 1000 dynes/sq cm. The pressure of the atmosphere at sea level is usually about 1000 mb. This pressure of 1000 mb is the same pressure which will support a column of mercury of density 13·5951 gm/cm^3 at a temperature of 0°C and subject to a standard gravity of 980·665 cm/sec^2, which is 29·530 in. (or 750·062 mm) in height. (*See* page 26.) (It will be remembered that 1 dyne is the unit of force (in the centimetre-gram-second system) which accelerates a unit mass (1 gram) by 1 cm/sec^2.) Barometers are now usually graduated in millibars and tables are available for converting readings in millibars to their equivalent values (in specified conditions discussed later) in inches or millimetres of mercury and vice versa. In temperate and high latitudes barometer readings at sea level usually lie between 970 and 1,030 mb, but readings as high as 1,050 mb and as low as 925 mb may occur occasionally, and substantially lower readings have been known near the centres of some tropical revolving storms. (A pressure of 1,000 mb is also known as 1 bar although this unit is not used in meteorology.)

If barometer readings are to be made comparable, the effect of the variation of the value of gravity between different latitudes, also of height above sea level and of temperature of the instrument must be eliminated. This is done by making small corrections, the process being known as the 'reduction of barometric pressure to standard conditions'. Details may be found in the *Marine Observer's Handbook*, but a brief summary is given here for the reader's convenience.

Barometer Corrections

Barometers graduated in millibars can be read to one-tenth of a millibar by means of a vernier. Mercurial barometers reading in millibars in use today, which were made before 1955, are calibrated so as to read as nearly as possible correctly at 12°C, in lat. 45°N at mean sea level. All barometers made or repaired on or after 1st January 1955, however, are made to read as nearly as possible correctly at 0°C and gravity 980·665 cm/sec² (which is practically the same as gravity at mean sea level in lat. 45°). In order to obtain the readings which the instrument would show under these conditions the following corrections must be applied (all references below to tables are to the Tables contained in the *Marine Observer's Handbook*):

(*a*) CORRECTION FOR TEMPERATURE (Tables VI and VII)

Mercury expands with increase of temperature, so that the same atmospheric pressure is balanced by a longer column of mercury on a warm day and by a shorter column on a cold day. A correction is applied to obtain the reading of the barometer which it would show at the standard temperature of 12°C or 0°C as the case may be.

(*b*) CORRECTION FOR ALTITUDE (Table X)

Atmospheric pressure, a measure of the weight of air above any point, will naturally be less at say the top of a lighthouse or on the bridge of a ship than at sea level some distance below. In order to make pressure readings all comparable, the pressure read on the instrument needs to be corrected to obtain the pressure which the instrument would record assuming it could be located at sea level in the same place. The actual difference depends not only upon the height of the instrument but also upon the density of the air between it and sea level. The density of the air in the column between the instrument and sea level varies according to the temperature of the air in that column and this is obtained from the reading of the dry-bulb thermometer in the screen.

(*c*) CORRECTION FOR LATITUDE (Tables VIII and IX)

Owing to the flattening of the earth at the poles and the fact that the vertical component of the centrifugal force due to the earth's rotation (which acts in the opposite direction from gravity) is greatest at the equator, the force of gravity increases steadily from the equator to the poles. The standard value of gravity is taken as 980·665 cm/sec², which is practically the same as that at sea level in lat. 45°. Thus a negative correction must be applied in low latitudes, where the barometer always reads high by a small amount because the 'weight' of the mercury is less than at lat. 45° (a positive correction must be applied in latitudes higher than 45°).

(*d*) INDEX CORRECTION

When all these corrections have been made it is found that the readings of individual barometers still differ slightly on account of differing values of capillarity of the mercury. These corrections are determined for every barometer at the National Physical Laboratory (NPL) and included in the 'index correction', the amount of which is stated on the certificate supplied by the NPL for use with the barometer.

EXAMPLE*

In lat. 27°N the barometer reads 1017·3 mb at a height of 53 ft above sea level. The attached thermometer reads 298°A, the dry bulb in the screen reads

* In the previous edition of *Meteorology for Mariners* the example given was for a mercurial barometer with a NPL certificate dated prior to 1955 (standard conditions 12°C, gravity at mean sea level in lat. 45°). The one here is for a barometer with a NPL certificate dated 1st January 1955 or later (standard conditions 0°C, gravity of 980·665 cm/sec²).

78°F and the index correction of the barometer is +0·3 mb. Standard conditions 0°C, gravity of 980·665 cm/sec.^{2}

		mb
Uncorrected reading		1017·3
Index correction..	+	0·3
		1017·6
Temperature correction for 298°A (25°C)	−	4·3
		1013·3
Height correction for 53 ft at air temperature of 78°F..	+	1·8
		1015·1
Gravity correction in lat. 27°	−	1·6
		1013·5

Marine mercurial barometers supplied by the Meteorological Office are fitted with a Gold Slide, which enables the corrections to be applied mechanically and dispenses with the use of tables. However, the corrections, as read off the Gold Slide, still need to be applied to the pressure reading!

Aneroid barometers need only to be corrected for index error and for altitude. These instruments work on the principle of the balancing of atmospheric pressure by a force due to the elastic deformation of a strong spring attached to a metal box from which the air has been exhausted and which is thereby prevented from collapsing under the external pressure of the atmosphere, and therefore they do not need to be corrected for changes in gravity. Aneroid barometers normally include a device for compensating for small changes of temperature, which is ensured either by using a bimetallic link or leaving a calculated small amount of air in the vacuum chamber. Aneroids compensated for temperature are usually so marked.

The readings of aneroid barometers after correction as above must be compared as frequently as possible with corrected readings of a good mercury barometer. The reason for this is that the index error of all aneroid barometers is liable to change quite frequently due to change in the elasticity of the metal of the vacuum chamber. (*See Marine Observer's Handbook*, Chapter 1.)

Pressure Distribution

Atmospheric pressure over the earth's surface is not constant but varies according to latitude and from place to place in the same general latitude. In the temperate zones, pressure is normally undergoing continual and often rapid change, while in the tropics it remains fairly constant for long periods, apart from a small diurnal range, which is primarily due to atmospheric tides. The distribution of land and water also influences pressure distribution, since these surfaces are subject to quite different temperature changes. (*See* Chapter 7.)

Isobars (Greek *Isos*, equal; *Baros*, weight)

If pressures, corrected as described above, are taken simultaneously at a number of different places and marked on a chart, a fair estimate can be made of those at intermediate places and lines can be drawn connecting places where the pressure is the same. The lines of equal pressure drawn on it are called ISOBARS and resemble the contours of equal altitude which define hills and valleys on a map. A chart for a particular time which includes data such as wind, weather, visibility and cloud and isobars (usually drawn at intervals of 4 mb, e.g. 1012, 1016 and 1020 mb, etc.) for a number of stations is termed a weather map, or more frequently a SYNOPTIC CHART, because it gives a synopsis or general view of the weather conditions over a large area at a given instant of time. Any synoptic chart will show a distribution of pressure in which there are regions of high pressure and low pressure resembling the hills, ridges and valleys found on a map. A typical region of high or low pressure on a weather map has a compact shape and may be 1,000 to 2,000 miles in width or diameter, and is seldom less than a few hundred miles in width. Before describing the main pressure features of the synoptic chart an understanding of the relation between pressure distribution and wind flow at the lowest levels is essential.

Pressure Gradient

The idea of pressure gradient is best illustrated by another look at a contour map where at any particular place we can estimate the gradient or slope of the ground. The steepest slope is seen to be at right-angles to the contour lines and its numerical value is greater where the contour lines are closer together. Similarly, we speak of the pressure gradient as being at right-angles to the isobars, its magnitude being measured by the ratio pressure difference/distance, in suitable units. Thus if two neighbouring isobars (4 mb interval) are 97 miles apart and are also straight and parallel, the pressure gradient is 4/97, i.e. 0·041 mb/mile between them, at right-angles to the isobars going from high to low pressure. A pressure gradient of this magnitude is associated with a wind which has a value of 30 knots in lat. 50°N or S. (*See* Chapter 3, page 35.) A complete definition of pressure gradient involves knowing its direction as well as its magnitude.

Pressure and Winds

A study of any synoptic chart of a large area outside the tropics, such as the North Atlantic, at once suggests that there is a close relation between pressure distribution and wind speed and direction; at any point the wind blows nearly parallel to the isobars in the neighbourhood, crossing them at a small angle in the direction from the higher to the lower pressure. The direction of the wind is such that to an observer facing the wind the lower pressure is on his right in the northern hemisphere and on his left in the southern hemisphere. This fact may be expressed in another way by saying that the air circulates clockwise round centres of high pressure and anticlockwise round centres of low pressure in the northern hemisphere (while in the southern hemisphere the air circulation is anticlockwise round centres of high pressure and clockwise round centres of low pressure). Lastly, the wind is strong where the pressure gradient is steep and light where the pressure gradient is flat, that is the speed of the wind is closely proportional to the pressure gradient.

All these facts were included in a law first stated by Buys Ballot in 1857 which will be given in Chapter 3.

The most important features in any pressure distribution are shown in Figs. 2.1–2.4 which are drawn for north latitudes.

(*a*) Depression or Low (Fig. 2.1)

A depression or low-pressure system is shown as a region of relatively low pressure with closed isobars. The isobars form a closed system with the lowest pressure inside that isobar which has the smallest value of pressure, the wind circulation around it being anticlockwise in the northern hemisphere, clockwise in the southern hemisphere. A description of the structure of a typical depression and the weather experienced within the various regions under its influence is given in Chapter 9, on pages 109 to 116.

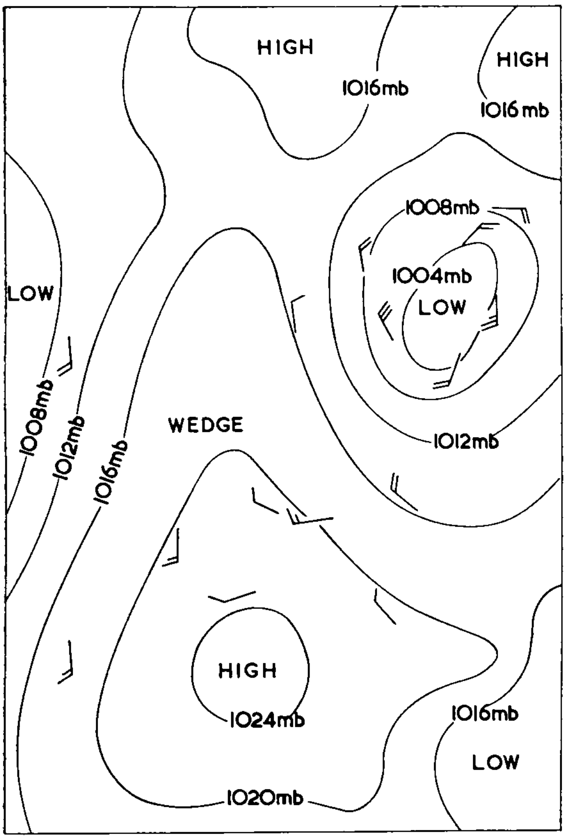


Fig. 2.1. Typical features of pressure distribution: anticyclone and depression

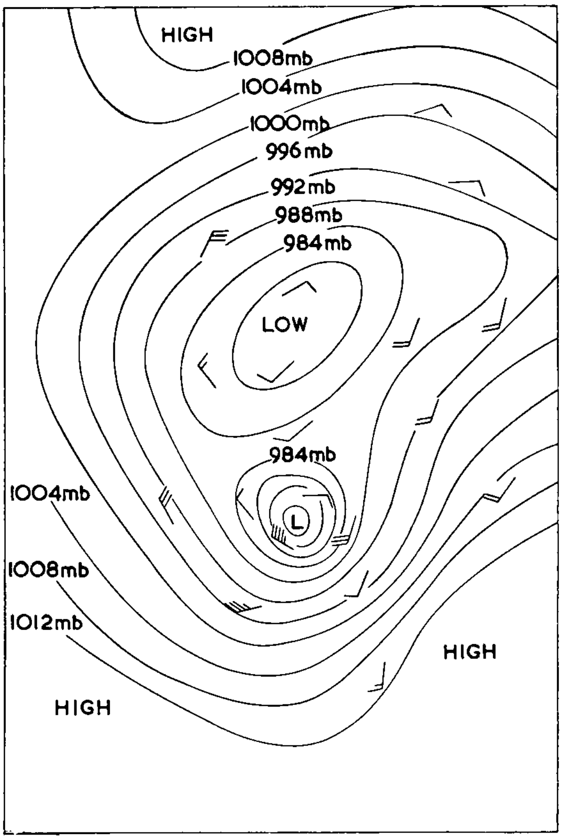


Fig. 2.2. Typical features of pressure distribution: primary and secondary depressions

(*b*) Secondary Depression (Fig. 2.2)

The term secondary depression is applied to a small depression within the area covered by a larger primary depression. The example shown in Fig. 2.2 is more vigorous than the parent or primary depression. Sometimes, however, no closed centre shows on the map, only a small region of relatively low pressure which maintains its identity for some time. The weather experienced in a secondary depression is described on page 118.

(*c*) Trough of Low Pressure (Fig. 2.3)

A trough is an elongated area of low pressure indicated by isobars extending outwards from a depression. In older text-books it was occasionally referred to as a V-shaped depression. This name is now obsolete. The weather associated with a trough of low pressure is summarised in Chapter 10, page 127.

(*d*) Anticyclone or High (Fig. 2.1)

An anticyclone is shown as a region of relatively high pressure with closed isobars. The isobars form a closed system with high pressure on the inside, the wind circulation around it being clockwise in the northern hemisphere and anticlockwise in the southern hemisphere. The weather experienced in an anticyclone is described in Chapter 10, page 124.

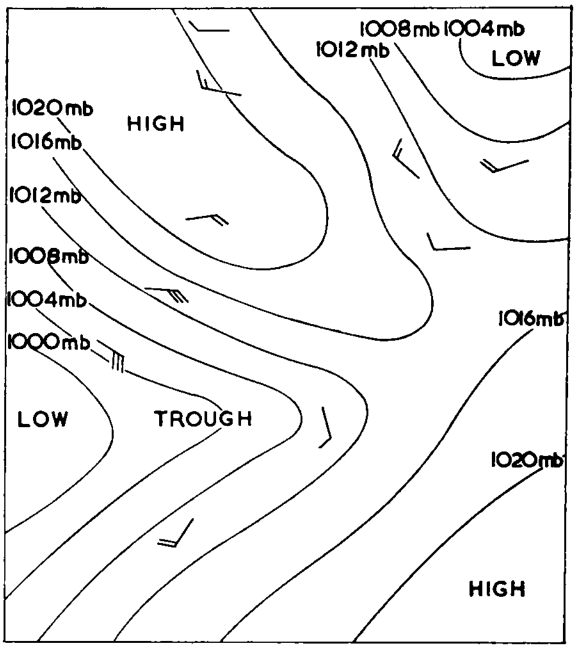


Fig. 2.3. Typical features of pressure distribution: depression and trough of low pressure

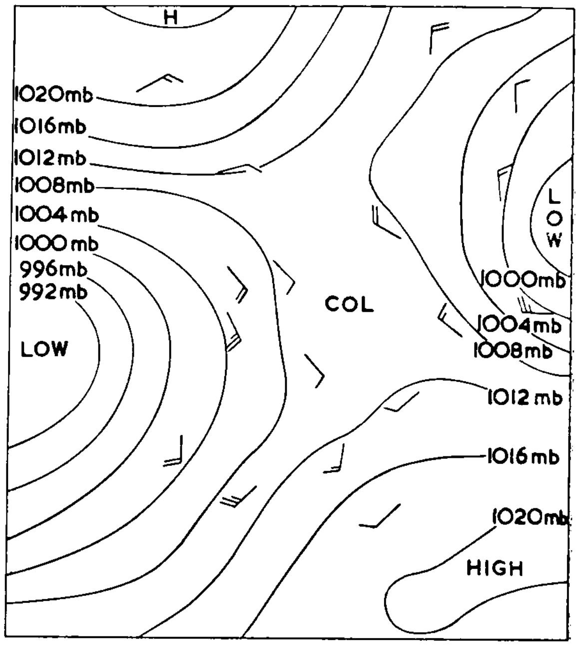


Fig. 2.4. Typical features of pressure distribution: col, depression and anticyclone

(*e*) Ridge of High Pressure or Wedge (Figs. 2.1 and 2.3)

A ridge of high pressure is an elongated area of high pressure indicated by isobars extending outwards from an anticyclone. The weather experienced in a ridge of high pressure is described in Chapter 10, page 127.

(*f*) Col (Fig. 2.4)

A col is a saddle-backed region between two lows and two highs. The weather experienced in a col is described on page 128.

The main features of a synoptic chart can be identified as belonging to these types of pressure distribution. An estimate of change in pressure distribution is a preliminary to the preparation of weather forecasts.

Diurnal Variation of Pressure

The frequently occurring changes of pressure as shown by a barograph trace may be due to many causes, and in temperate latitudes in winter it is not possible to discern any systematic variation. But in the tropics, and in quiet summer weather of the temperate zones, a diurnal variation of pressure with a definite pattern becomes evident. Over a long period the mean daily range is less than 1 mb in the British Isles but more than 2 mb in subtropical and tropical regions. Fig. 2.5 shows a typical curve for the tropics, with maxima at 1000 and 2200 hours and minima at 0400 and 1600 hours local time.

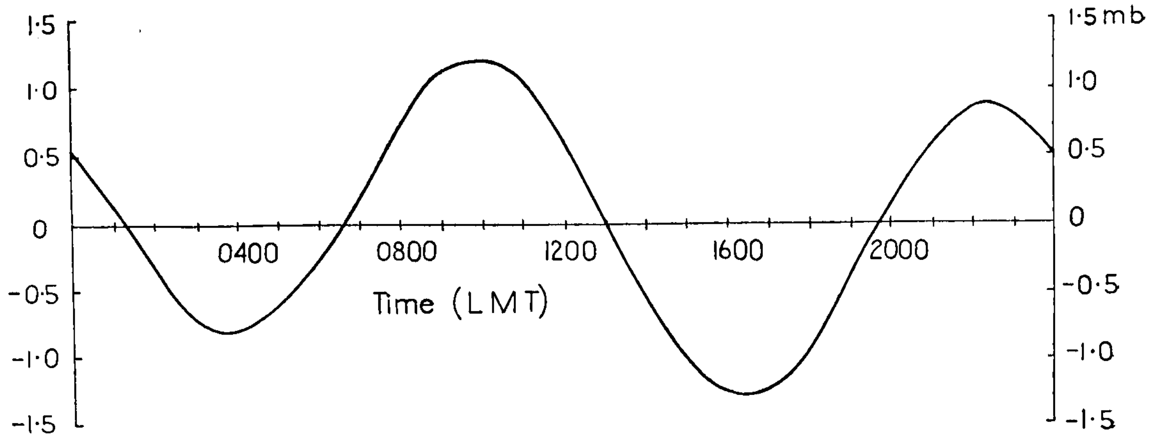


Fig. 2.5. Diurnal variation of pressure typical of the tropics

In temperate latitudes irregular pressure changes are usually so much larger than the diurnal variations that the latter need not be taken into account. In tropical regions irregular changes are usually much smaller than the diurnal change from which they cannot be distinguished until the diurnal change has been subtracted. This procedure is important in tropical regions because a fall of pressure of 2 or 3 mb below the value appropriate for the locality, after allowing for the regular diurnal change of pressure, may be the first indication of the approach of a tropical revolving storm. The normal value of pressure for a locality and the diurnal change can be obtained from a meteorological atlas or *Admiralty Pilot.*

Pressure Tendency

The name TENDENCY is given to describe the rate of change of pressure with time. In practice it usually refers to the time interval of three hours before an observation. The meteorologist wishes to know the change of pressure at a particular place. The ship's report of barometric tendency does not directly tell him this, because part of that tendency is due to the ship's progressive movement. But by using the figures for the ship's course and speed given in the synoptic message he can eliminate this spurious part of the tendency by making a correction to the reading.

Corrected readings of barometric tendency help to determine the movements of the various pressure systems, since the barometer usually falls in advance of a depression and rises in advance of an anticyclone. If the system is stationary, the tendency can show whether it is filling up or deepening. (*See* Chapter 12.)

Lines drawn on the chart through places having the same tendency are known as ISALLOBARS. The isallobar is of value to the forecaster because it shows a rate of change of pressure with respect to time.

CHAPTER 3

WIND AND WAVES

Introduction

Wind is defined as air in motion. In general the wind motion estimated by an observer or measured by an anemometer is the wind parallel to the ground. Ashore, wind is measured by anemometers which record its velocity in knots. As the wind speed near the surface varies with height above the ground, readings of anemometers are corrected to a standard height of 33 ft (10 metres) to make accurate comparisons possible between readings of these instruments when sited at different levels. Except in naval ships, anemometers are rarely used at sea owing to expense and siting difficulties and because the ship's structure complicates the air flow and makes accurate readings hard to obtain. At sea, observations of wind are therefore normally made by estimation.

Beaufort Scale

The BEAUFORT SCALE forms the basis of wind-force estimation at sea. It was originally introduced in 1808 by Admiral Beaufort, who defined the numbers of the scale in terms of the effect of wind on a man-of-war of his day. Table 3.1 shows how the method has changed through the years. The scale itself has not changed much, however, and it seems likely that conditions which Beaufort would have described by a certain number, judged from the behaviour of a warship of his time, would be described by the same number using the modern criterion whereby the wind force is judged from the appearance of the sea. As will be seen from the table each Beaufort number has been allotted an equivalent range of wind speeds as well as an equivalent mean wind speed at standard height, so that conversion from Beaufort number to wind speed is simple. Wind direction at sea is also normally estimated from the appearance of the sea—the true azimuth from which the wind is blowing being that at right-angles to the line of waves. For further details about wind observation *see Marine Observer's Handbook*, Chapter 4.

Wind and Air Movements on a Rotating Earth

Wind is movement of air set up in the atmosphere by differences in atmospheric pressure between two localities. These differences in pressure are caused by variations of temperatures in columns of air over different places. The atmosphere is always trying to achieve a uniform pressure distribution by transfer of air from one region where an accumulated excess of air has resulted in high pressure, to another region where a deficiency of air has resulted in low pressure. It is a matter of common observation that this adjustment is not carried out by winds blowing direct from high to low pressure but that the wind tends to blow in a circular manner around regions of low pressure or high pressure (*see* page 28), and for the explanation of this we must consider the effect of the rotation of the earth.

At the equator the velocity of the earth's surface is about 1,000 miles per hour from west to east, gradually decreasing towards the poles where it is zero. An

Table 3.1. Beaufort Scale of Wind Force

Beaufort Number	General Description	Beaufort's Criterion	Sea Criterion	Landsman's Criterion	Limits of Velocity in knots (Measured at a height of 33 ft above sea level)	Average Velocity in knots (Measured at a height of 33 ft above sea level)	Equivalent pressure in pounds per square foot
0	Calm	Calm.	Sea like a mirror.	Calm; smoke rises vertically.	Less than 1	0	0
1	Light air	Just sufficient to give steerage way.	Ripples with the appearance of scales are formed without foam crests.	Direction of wind shown by smoke drift, but not by wind vanes.	1 to 3	2	0·01
2	Light breeze	That in which a well-conditioned man-of-war with all sail set and 'clean full' would go in smooth water from 1 to 2 knots.	Small wavelets, still short but more pronounced. Crests have a glassy appearance and do not break.	Wind felt on face; leaves rustle; ordinary vane moved by wind.	4 to 6	5	0·08
3	Gentle breeze	3 to 4 knots.	Large wavelets. Crests begin to break, Foam of glassy appearance. Perhaps scattered white horses.	Leaves and small twigs in constant motion. Wind extends light flags.	7 to 10	9	0·28
4	Moderate breeze	5 to 6 knots.	Small waves, becoming longer; fairly frequent white horses.	Raises dust and loose paper; small branches are moved.	11 to 16	13	0·67
5	Fresh breeze	That to which she could just carry in chase 'full and by' Royals, etc.	Moderate waves, taking a more pronounced long form; many white horses are formed. Chance of some spray.	Small trees in leaf begin to sway. Crested wavelets form on inland waters.	17 to 21	18	1·31
6	Strong breeze	Single-reefed topsails and top-gallant sails.	Large waves begin to form; the white foam crests are more extensive everywhere. Probably some spray.	Large branches in motion; whistling heard in telegraph wires, umbrellas used with difficulty.	22 to 27	24	2·3
7	Near gale*	Double-reefed topsails, jib, etc.	Sea heaps up and white foam from breaking waves begins to be blown in streaks along the direction of the wind.	Whole trees in motion; inconvenience felt when walking against wind.	28 to 33	30	3·6
8	Gale*	Triple-reefed topsails, etc.	Moderately high waves of greater length; edges of crests begin to break into the spindrift. The foam is blown in well-marked streaks along the direction of the wind.	Breaks twigs off trees; generally impedes progress.	34 to 40	37	5·4
9	Strong gale	Close-reefed topsails and courses.	High waves. Dense streaks of foam along the direction of the wind. Crests of waves begin to topple, tumble and roll over. Spray may affect visibility.	Slight structural damage occurs (chimney-pots and slates removed).	41 to 47	44	7·7
10	Storm*	That which she could scarcely bear with close-reefed main topsail and reefed foresail.	Very high waves with long overhanging crests. The resulting foam in great patches is blown in dense white streaks along the direction of the wind. On the whole the surface takes on a white appearance. The tumbling of the sea becomes heavy and shock-like. Visibility affected.	Seldom experienced inland; trees uprooted; considerable structural damage occurs.	48 to 55	52	10·5
11	Violent storm*	That which would reduce her to storm staysails.	Exceptionally high waves (small and medium-sized ships might be for a time lost to view behind waves). The sea is completely covered with long white patches of foam lying along the direction of the wind. Everywhere the edges of the wave crests are blown into froth. Visibility affected.	Very rarely experienced; accompanied by wide-spread damage.	56 to 63	60	14·0
12	Hurricane	That which no canvas could withstand.	The air is filled with foam and spray, Sea completely white with driving spray; visibility very seriously affected.		64 and over	—	—
		Devised by Admiral Beaufort and brought into use in 1808.	Agreed to by International Meteorological Committee in 1939, amended in 1941 and further amended by World Meteorological Organization in 1953 and 1958.	Devised in 1906 by Dr. G. C. Simpson, later Director of the Meteorological Office.			

* Up till 1958, when amended as above by the World Meteorological Organization, the general description for Beaufort numbers 7 to 11 were:
7—Moderate gale; 8—Fresh gale; 9—Strong gale; 10—Whole gale; 11—Storm.

atmosphere which was completely calm relative to the earth's surface would still share this rotation which is relative to space. To understand the effect of the earth's rotation on the atmosphere it is simplest to consider first what happens to a ring of air round the earth at the equator which is displaced northwards. As a result its radius would be reduced and its absolute speed of rotation from west to east would be increased, so that the air in the ring would become a belt of west wind relative to the earth (*see* Fig. 3.1). This occurs on account of a principle known as the 'conservation of momentum', which can be illustrated by attaching a marble to a string and swinging it around the free end. When the length of the string is halved the angular velocity of the marble becomes four times its original value. A similar argument shows that a ring of air in any latitude which was displaced towards the equator would suffer a decreased speed of rotation relative to the earth because its radius of rotation would be increased and hence the air in this ring would become a belt of east wind over the surface of the earth.

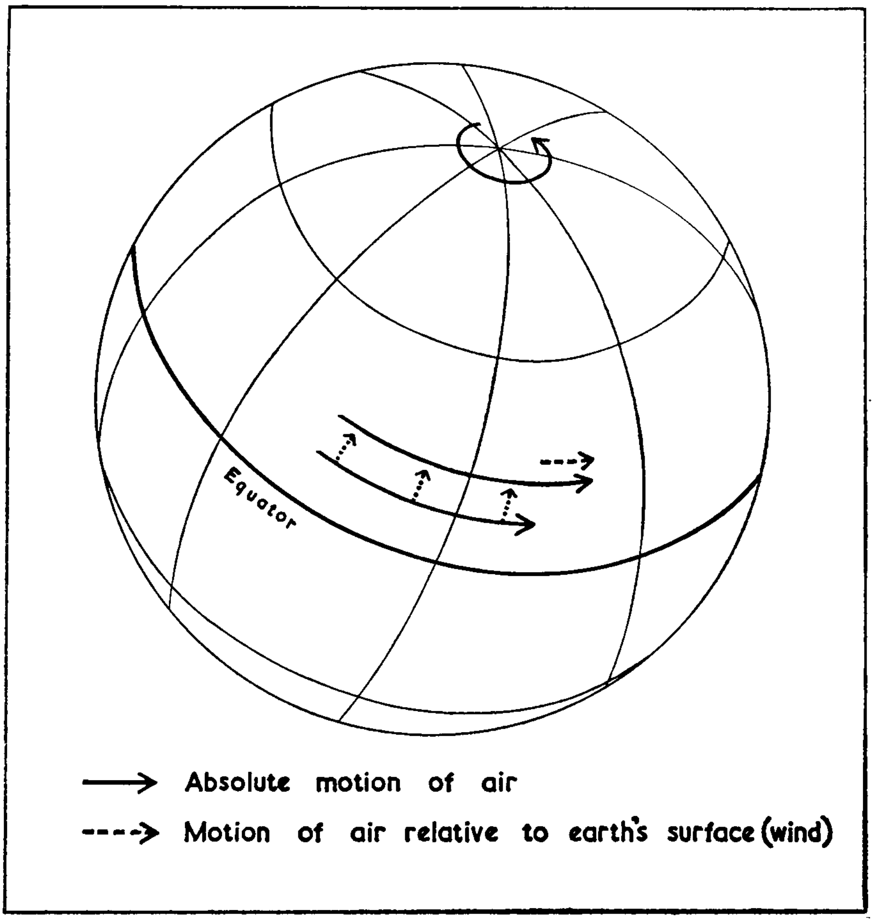


Fig. 3.1. West wind caused by poleward displacement of air

If it were not for this effect due to rotation, the wind distribution over the world would be relatively simple, because air strongly heated in tropical regions would ascend by convection to the upper troposphere, resulting in winds blowing towards the equator at low levels in most latitudes in both northern and southern hemispheres, while at higher levels winds would blow towards the polar regions. Since the rotation of the earth and the conservation of momentum are facts, it follows that W winds tend to be established in latitudinal belts of air which are displaced polewards, and E winds tend to be established in similar belts of air which are displaced towards the equator; so for this reason alone the simple picture of N–S air currents just described does not fit the facts.

Now let us consider again the case of a ring of air right round the earth which has recently been displaced away from the equator. On account of the principle of conservation of angular momentum we see that it will acquire an additional velocity from west to east (this is true in both hemispheres), hence it will behave as a belt of westerly wind, while the air on both sides of it is at rest relative to the earth's surface. (In practice turbulence would not allow this idealised situation to persist for more than a short time, but the purpose of this illustration is to explain the principal forces at work.) Since this belt of air will then be rotating around the earth's axis with a greater angular velocity than the air on either side of it, it will be acted on by a centrifugal force greater than that which acts on the adjacent air which is at rest relative to the earth's surface. This excess centrifugal force will tend to throw the whole ring away from the earth's axis.

At the poles the whole of this centrifugal force will be towards the equator, and at the equator the whole force will be upwards and away from the earth's surface. At any other latitude the force can thus be resolved into its two components—but for surface wind movement we need only consider the equator-wards force, which varies in magnitude with the latitude.

To keep a 'westerly' wind belt from being thus thrown equator-wards in the northern hemisphere the atmospheric pressure must be higher to S than to N, thus producing a pressure force (gradient) capable of balancing the excess centrifugal force. Putting it another way, if this pressure force (gradient) is not available, then the whole ring of air will be displaced to the southwards until enough air has piled up on its equator-ward side to bring about equilibrium.

The relationship between wind and pressure gradient is thus mutual. At every point in a particular pressure distribution the air will be moving in such a way that an observer facing the wind in the northern hemisphere will find that the air pressure will fall from left to right across the wind stream, and the strength of the wind will be directly proportional to the magnitude of the pressure gradient. These facts were summarized by Buys Ballot in 1857 in a law now known by his name.

Buys Ballot's Law

In north latitudes, face the wind and the barometer will be lowest to your right.

In south latitudes, face the wind and the barometer will be lowest to your left.

Wind Circulation around Pressure Systems

The first deduction which can be drawn from this law is that the winds circulate around the centres of low and high pressure; in the northern hemisphere they blow anticlockwise around an area of low pressure and clockwise around an area of high pressure, while in the southern hemisphere the directions are reversed. A look at any synoptic chart will verify that this statement is nearly true except for regions within a few degrees of latitude of the equator, although it will be noticed that over land the wind blows mostly about 30° from the direction of the isobars towards the side of low pressure, and over the sea the inclination of the wind to the direction of the isobars is about 10°, but is usually greater than this in the Trades. In both cases this effect is mainly caused by friction due to the irregularity of the land or sea surface, which also retards the air in the lowest layers. At a level of about 2,000 ft and higher the wind blows parallel to the isobars in agreement with Buys Ballot's Law.

Relation between Wind and Pressure Gradient

To relate wind and pressure gradient, we should imagine a small parcel of air to be isolated and determine all the forces that can set it in motion. At first sight, these are the weight acting vertically downwards and the pressure acting horizontally along the pressure gradient from high to low pressure. A vertical force has no effect in creating horizontal motion, so we are left with the horizontal pressure alone. If there were no rotation of the earth the air would thus tend to move along the pressure gradient (i.e. *across* the isobars) direct from high to low pressure.

This nearly fits the motion of air within a few degrees of latitude of the equator in a region where the pressure gradient and the changes of pressure are small, but elsewhere it does not agree with the fact that, as shown by Buys Ballot's law, the wind blows *along* the isobars with only a small deviation towards the side of low pressure. The discrepancy is due to our having ignored the effect of the earth's rotation and of friction.

Effect of the Earth's Rotation

Let us imagine an observer at any point **P** on the earth's surface (northern hemisphere, lat. φ). The earth's rotation around its axis **OA** may be represented by the angular velocity ω (*see* Fig. 3.2). From the point of view of our observer it can equally well be represented by an angular velocity ω around **PD** parallel to **OA**. Now the angular velocity about any point on a rotating sphere may be split up into two components around axes at right-angles to one another, in the same way as a force F can be split up into components along two perpendicular axes. So the earth's angular velocity about **PD** can be resolved into two components round **PC** and **PB** and represented by:

$$\omega \cos \varphi \quad \text{about } \mathbf{PC},$$
$$\omega \sin \varphi \quad \text{about } \mathbf{PB}.$$

As we are solely concerned with horizontal motion, which is only affected by $\omega \sin \varphi$ (the component of angular velocity about the vertical line **PB** at right-angles to the earth's surface at **P**), $\omega \cos \varphi$ (the component of angular velocity about the line **PC** in the plane of the earth's surface) can be neglected here.

Fig. 3.3 is drawn in the tangent plane of **P** in Fig. 3.2, which we have just explained is rotating (anticlockwise as shown here) about **PB** with angular velocity $\omega \sin \varphi$ due to the earth's rotation. In this case **PB** is perpendicular to the paper. Suppose a particle moving from **P** towards **E** with velocity V: in a time t, it will travel a distance Vt. In the same time the earth's surface, i.e. the tangent plane at **P**, will turn through an angle $\omega t \sin \varphi$ so that relative to the earth's surface the particle reaches **F** instead of **E**. The distance **EF** is $Vt\,\omega t \sin \varphi$ ($= Vt^2\omega \sin \varphi$) so long as t is small.

The distance **EF** thus depends on the square of the time in which the air has been moving from **P**. But in simple dynamics we find that an acceleration f acting on a body for time t moves it a distance in accordance with the formula $s = \frac{1}{2}ft^2$. Thus the particle is subject to an acceleration f whose value we can

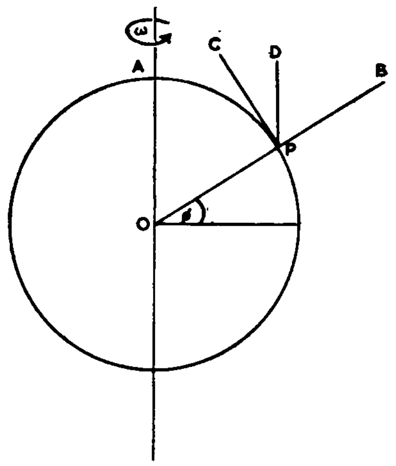


Fig. 3.2. The effect of the earth's rotation

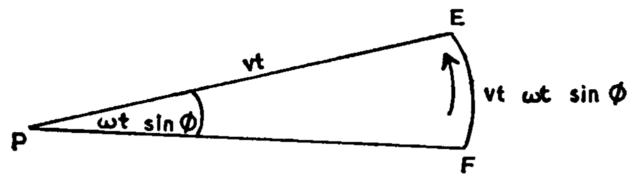


Fig. 3.3. Air movement on a rotating earth

determine by writing **EF** for the distance s and hence $\frac{1}{2}ft^2 = \textbf{EF} = Vt^2\,\omega\sin\varphi$ whence $f = 2\omega\,V\sin\varphi$.

Thus every particle moving with a velocity V horizontally over the earth's surface is subject to an acceleration of magnitude $2\omega\,V\sin\varphi$ perpendicular to the instantaneous direction of motion of the particle. This acceleration is due to the earth's rotation.

Since the magnitude of this acceleration depends upon $\sin\varphi$, it is smallest within a few degrees of the equator where the air movement (wind) is as a result often down the pressure gradient, that is *across* the isobars, whereas in high latitudes where the acceleration has a larger value, increasing to $2\omega V$ at the poles, the wind blows more nearly in agreement with Buys Ballot's law, that is *along* the isobars.

Geostrophic Wind

We shall now consider air motion in a region far enough from the equator to make the earth's rotation effect important. If the wind is blowing steadily along a straight line (i.e. in a region where the isobars are straight and parallel, as shown in Fig. 3.4) then, provided we neglect frictional forces, there are only two horizontal forces acting on any parcel of air: one due to the pressure gradient

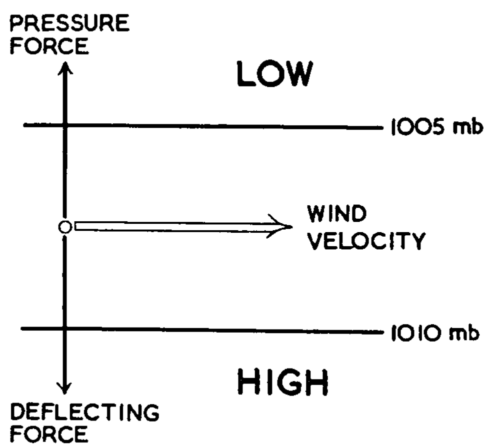


Fig. 3.4. Geostrophic wind velocity

acting in a direction perpendicular to the isobars towards the side of low pressure, the other the deflecting force of the earth's rotation acting perpendicular to the path of the air and to the right of it. Only if the accelerations produced by these two forces balance, i.e. are equal in magnitude and opposite in direction, can the motion remain steady and V remain constant as was assumed above. Denoting the pressure gradient by G and the density of the air by ρ, the acceleration due to the pressure gradient is given by $\frac{G}{\rho}$. For steady motion this acceleration must balance the acceleration $2\omega V \sin \varphi$ due to the earth's rotation, i.e.

$$\frac{G}{\rho} = 2\omega V. \sin \varphi \text{ or } V = \frac{1}{2\omega \sin \varphi} \cdot \frac{G}{\rho}.$$

The wind velocity V derived on the assumptions that the isobars are straight and parallel, and ignoring friction, is called the GEOSTROPHIC WIND. Its direction is parallel to the isobars, with low pressure to the left in the northern hemisphere and to the right in the southern hemisphere.

Modifying Effect of Friction

When the isobars are straight and parallel, the geostrophic wind derived from the surface pressure distribution has been found to be a good approximation, in the temperate zone, to the wind at about 2,000 ft, i.e. high enough above the surface to be unaffected by friction. The surface wind, however, does not blow precisely along the isobars in any latitude but at an angle to the isobars to the side of low pressure. It can be shown that friction will produce exactly this effect; the actual amount of deviation from the isobars depends not only on the latitude but also on the amount of friction.

Another effect of surface friction is that the wind speed at the surface is reduced below the value of the geostrophic wind. Over the sea, it is found that the surface wind speed usually is about two-thirds that of the geostrophic wind. Over land, the fraction varies more widely and generally lies between two-thirds and one-half of the geostrophic wind speed, according to the degree of exposure of the place where the wind is measured.

Determination of Surface Wind from Isobars

The equation derived above allows the geostrophic wind to be calculated from the pressure gradient and shows that for a given pressure gradient the geostrophic wind has the same value along a given latitude. Synoptic weather maps contain a diagram from which geostrophic wind speed can be determined by measuring the distance between successive isobars. If this diagram, on which the various wind speeds are given in knots, is transferred to a celluloid scale, the geostrophic wind at any locality on the synoptic chart can be read off by simply putting the scale across the isobars as shown in Fig. 3.5. In the example shown, where the scale is constructed for isobars drawn at intervals of 4 mb, the geostrophic wind between the isobars of 996 and 1,000 mb is found to be 11 knots. A correction for latitude can be applied where necessary.

The speed of the surface wind at sea may then be taken as two-thirds of the geostrophic wind value, and the direction as one or two points from the direction of the isobars, towards the side of low pressure.

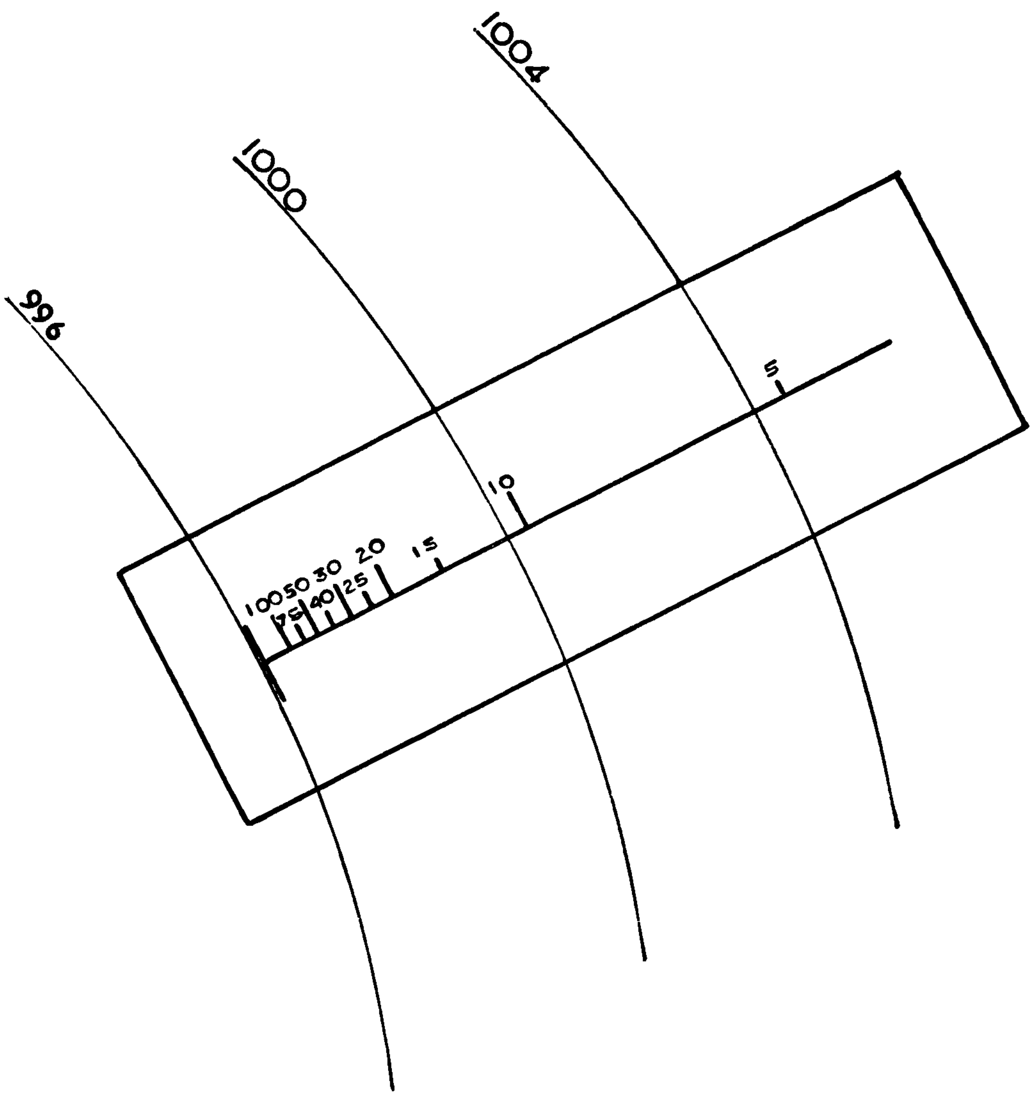


Fig. 3.5. The determination of surface wind from isobars

Gradient Wind

It has already been said that the geostrophic wind is only a good approximation to the true wind above the 'friction layer' (up to 2,000 ft) when the isobars are straight and parallel. The wind which blows along curved isobars is known as the GRADIENT WIND. In this case, the direction of motion approximates to the direction of the isobars if there is no friction and motion is steady. Figs. 3.6 and 3.7, drawn for the northern hemisphere, show two examples of curved isobars, the first representing cyclonic and the second anticyclonic motion. Arrows with a double shaft denote velocity V, while single arrows denote the forces which affect the air at **A** and **B**. For steady circular motion, the difference between the accelerations due to the pressure gradient and the deflecting force must be exactly that required to keep the air moving in a circular path. From the elementary formula for centrifugal force $\left(\frac{mV^2}{r}\right)$, where m is the mass of air and r is the radius of curvature of the isobars, this acceleration can be shown to be $\frac{V^2}{r}$. A common example of centrifugal force is the tendency for a weight

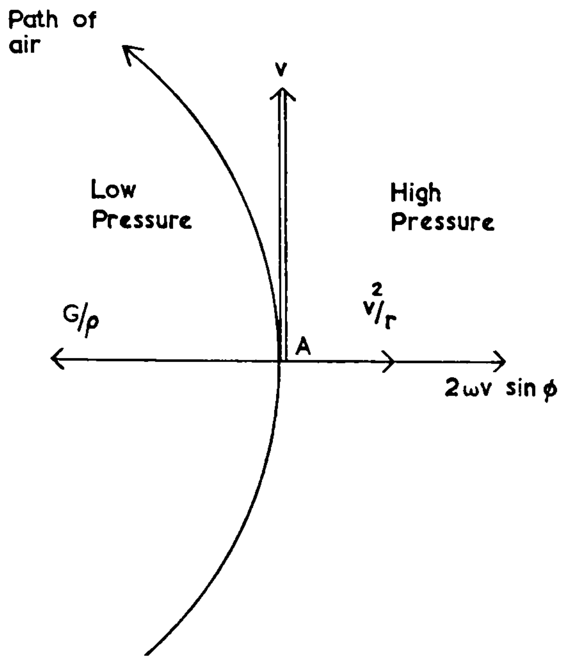


Fig. 3.6. Gradient wind for cyclonic motion

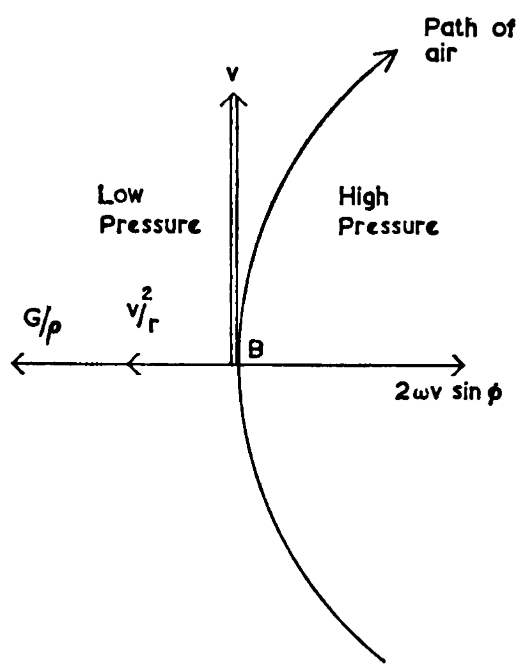


Fig. 3.7. Gradient wind for anticyclonic motion

to fly off into the air if whirled round on a piece of string. The centrifugal force always acts outwards, at right angles to the tangent to the curved path along which the mass is moving. Thus for a cyclone the centrifugal force is directed against and has the opposite sign to the pressure gradient, while for an anticyclone it acts in the same direction and has the same sign as the pressure gradient. The signs of the term $\frac{V^2}{r}$ are as shown in equations (1) and (2). We know that acceleration due to the pressure gradient $= \frac{G}{\rho} = 2\omega V \sin \varphi$ and we have therefore, for cyclonic motion (Fig. 3.6)

$$\frac{G}{\rho} - \frac{V^2}{r} = 2\omega V \sin \varphi. \qquad (1)$$

and for anticyclonic motion (Fig. 3.7)

$$\frac{G}{\rho} + \frac{V^2}{r} = 2\omega V \sin \varphi. \qquad (2)$$

Each of these quadratic equations can give two values for V but it can be shown that one of these values does not represent real motion.*

If we now transpose equation (1) into the form

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

it can be seen that, for cyclonic motion, the gradient wind speed is less than geostrophic wind by the quantity

$$\frac{V^2}{r 2\omega \sin \varphi}.$$

* On a very much smaller scale, however, clockwise depressions in the northern hemisphere can occur, e.g. in dust devils, where the initial clockwise rotation might be caused mechanically, by turbulence or otherwise, and where 2ω V sin φ is very small compared to the other terms so that the pressure gradient and the centrifugal force are in balance.

Similarly, transposing equation (2) into the form

$$V = \frac{G}{\rho 2\omega \sin \varphi} + \frac{V^2}{r2\omega \sin \varphi}$$

we see that, for anticyclonic motion, the gradient wind speed is greater than geostrophic wind by the quantity

$$\frac{V^2}{r2\omega \sin \varphi}.$$

Table 3.2 at page 48 illustrates the difference between geostrophic and gradient wind speed for various radii of curvature of the isobars, in lat. 55°.

Cyclostrophic Wind

In the equation for cyclonic motion:

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

the term on the right-hand side, which represents the acceleration due to the deflecting force of the earth's rotation, is evidently small in comparison with the terms on the left-hand side of the equation when G (pressure gradient force) is large and r (radius of storm) is small. This state of affairs corresponds to those found in a tropical revolving storm and even more closely to the conditions occurring in a tornado or a waterspout. In the case of the tropical revolving storm, which spends most of its life in low latitudes, the right-hand term is also small because φ is a small angle and therefore sin φ has a small value. Hence we can write the equation as:

$$\frac{G}{\rho} - \frac{V^2}{r} = 0$$

giving

$$V = \sqrt{\frac{Gr}{\rho}}.$$

This value of V is known as the CYCLOSTROPHIC WIND and represents the wind which is found to blow nearly parallel with the (more or less) circular and crowded isobars in a tropical revolving storm.

Although the cyclostrophic wind does in fact give a good approximation to the actual wind found in most tropical revolving storms and tornadoes, nevertheless the effect of the earth's rotation does have some influence on the behaviour of these systems. For instance, tropical revolving storms seldom, if ever, form within a few degrees of latitude of the equator (usually about 5°) and, when formed, the winds in them blow around their low-pressure centres in the direction which is appropriate to the hemisphere, that is anticlockwise in the northern hemisphere and clockwise in the southern hemisphere. Although the origin of tropical revolving storms is not yet fully understood, there is now little doubt that they begin their existence at the centre of a small area of low pressure. The fact that they do not seem able to develop really intense wind circulation in the neighbourhood of the equator is probably due to the absence of the effect due to the earth's rotation. As soon as a cyclonic wind circulation can be recognised it is found to resemble the circulation normally found around a low-pressure area in the appropriate hemisphere, throughout the subsequent life of the storm. Thus, although the cyclostrophic wind gives a useful approximation to the winds found in a tropical revolving storm, it is important to bear in mind that the deflecting force due to the earth's rotation has a decisive influence on the behaviour of the storm at all stages in its life history.

Wind Structure at the Lowest Levels

It is a matter of everyday observation that the speed and direction of the wind are constantly varying over short periods. A knowledge of the main features of these variations is very important for the meteorologist, since they have been found to be closely related to the turbulence and lapse rate of the atmosphere as well as to the roughness of the surface, whether land or sea. For the past half-century wind structure near the surface has been thoroughly examined with the aid of daily records made by pressure-tube anemographs at numerous inland, coastal and island reporting stations in the British Isles and other countries. An example of a typical record from this instrument is shown in Fig. 3.8, from which it can be seen that the open scale used enables the magnitude of individual gusts to be measured and the variations of wind in periods of a few minutes, e.g. in squalls and thunderstorms, to be studied in detail.

The main feature of the wind structure is that although both speed and direction often maintain the same average values during a period of, say, several successive hours, nevertheless from minute to minute both speed and direction show a considerable range, the effect of which is to broaden the traces of both speed and direction in the characteristic way shown by almost any anemograph record. A typical example is shown in Fig. 3.8. In terms of the wind itself there is a continuous succession of gusts and lulls associated with equally rapid changes of direction over a range which may exceed 30°.

This effect is most marked over land, where the gustiness, as shown by the magnitude of the variations of wind speed and direction between successive gusts and lulls in the wind during a period of a few minutes, is much greater than over the sea. In terms of the anemograph record the typical trace from a station with an exposure like that of the Weather Centre roof in London shows a broad ribbon for both speed and direction, whereas over the sea at a typical station such as the Bell Rock Lighthouse, 12 miles off the east coast of Scotland, the trace is quite narrow, showing a small difference between gusts and lulls. In the case of Bell Rock there is a clear difference between winds from different directions, winds from the west having travelled mainly over land are noticeably more gusty than east winds, which have travelled a long distance over the North Sea.

Thus the wind is seldom steady but consists of a series of gusts and lulls at short intervals, and its speed and direction vary. Short-period wind variations, such as are found in any anemograph trace, are partly due to eddies caused by the roughness of the surface over which the wind travels. The mean wind speed over a period of time is therefore the mean of many gusts and lulls. Anemograph traces of wind at sea have a much narrower band of variations both of speed and direction, showing that the 'roughness' of the sea surface is much less than that of the land.

A gust is any sudden increase of wind of very short duration. A measure of the intensity of a gust is given by the peak velocity, which is easily recognizable on the trace of a pressure-tube anemometer.

A squall comprises a rather sudden increase of the mean wind speed which

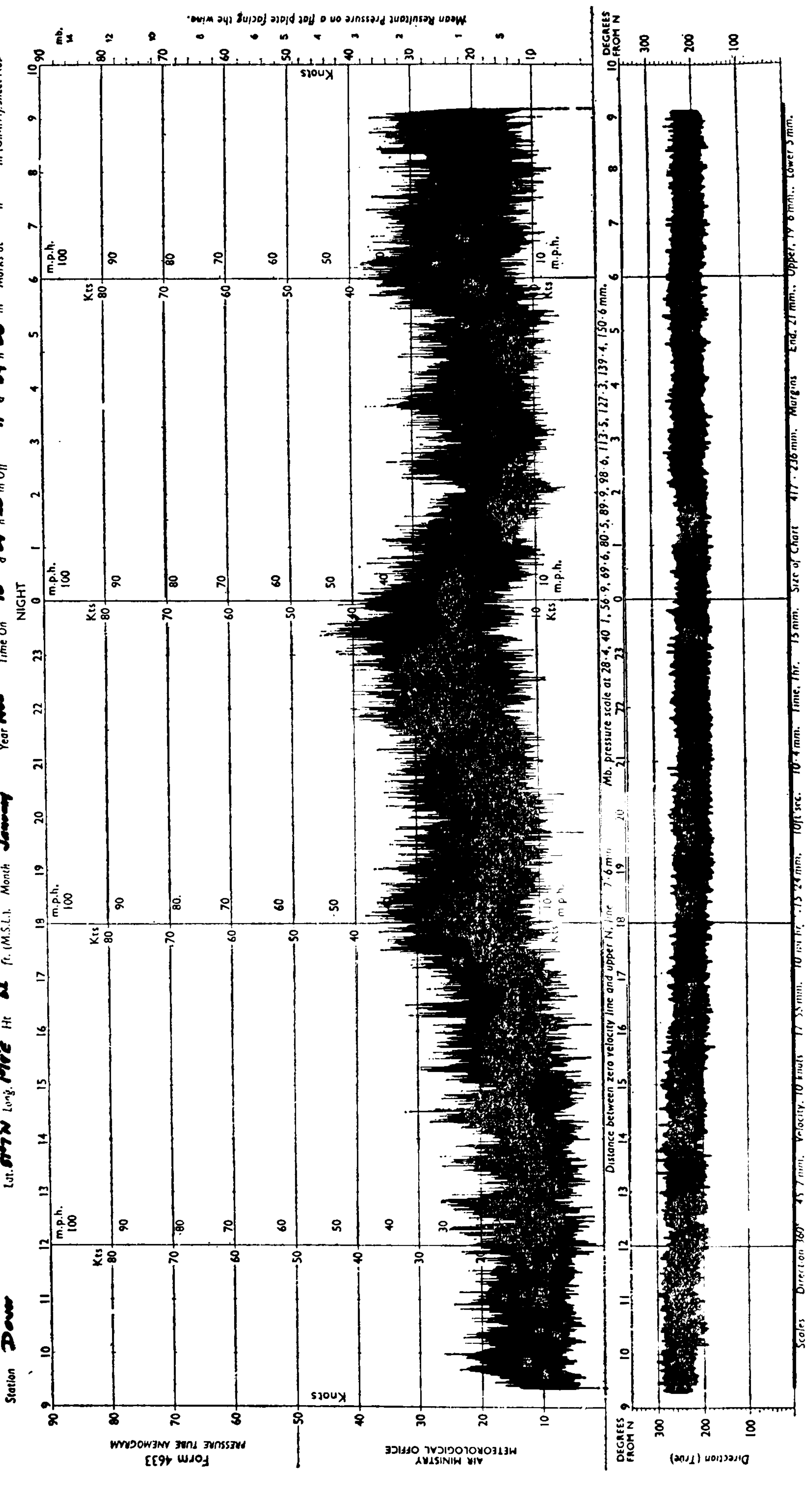


Fig. 3.8. Wind speed and direction as recorded by a pressure-tube anemograph

lasts for several minutes at least, before the mean wind speed returns to near its previous value. A squall may include many gusts.

The cause of the gustiness of the wind also lies in the formation of eddies due to turbulence. In a very light breeze, wind speed and direction over a smooth surface are sometimes nearly constant, because the air flows quite smoothly. A small increase in wind speed is sufficient to break up the flow into eddies, or whirls of irregular shape and size which resemble in many ways the eddies visible on the surface of a fast-flowing river or in any river just below the piers of a substantial bridge. Through the agency of turbulence inequalities in the air are smoothed out faster than would otherwise be the case.

The increase of wind with height which is most pronounced near the earth's surface has been described earlier. It can equally well be looked on as a reduction of the wind speed below the geostrophic value from about 2,000 ft to the surface, and which becomes most marked in the lowest layers. Irregularities in the land or sea surface bring about this reduction of wind speed, by causing turbulent eddies to form within the air flow near the surface. These eddies consume energy and thereby reduce the wind speed, making it seem as though the land or sea surface is exerting a frictional drag on the air flow near the ground. Over hilly country, forests or built-up areas many more eddies are produced and the gustiness and reduction of the wind speed below the geostrophic value is therefore much greater than over open level country and the open sea.

Within the layer affected by friction the wind direction veers with height in the northern hemisphere from the surface upwards until near 2,000 ft it blows parallel to the isobars.

It was stated at the beginning of this chapter that wind structure in the lowest layers was related to the lapse rate as well as to turbulence. The reason for this can now be appreciated. Consider first occasions when the lapse rate is large, i.e. steeper than the dry adiabatic, as often occurs when strong winds of polar origin are carrying air rapidly towards warmer latitudes and the air is much colder than the sea. Such conditions favour vigorous convection, which ensures that turbulent eddies formed in the strong wind prevailing are carried upwards rapidly. Turbulence is also responsible for bringing other eddies from a higher level downwards to a lower one. In the absence of turbulence and convection, the frictional effect of the sea surface would ensure that the horizontal velocity of the air at the surface would be less than that of the air at 2,000 ft, i.e. the momentum of eddies near the surface would be less than that of eddies at 2,000 ft. However, in the case considered, surface layers are being rapidly mixed with other layers at higher levels; the sharing of momentum thus effected ensures that the mean surface wind over a short period will differ little from the geostrophic wind at about 2,000 ft in speed and direction, although its gustiness is rather accentuated.

In the opposite case with a small lapse rate or an inversion in the lowest layers, say below 2,000 ft, which is associated with warm air moving over a relatively cold sea surface, convection currents cannot develop so that turbulent eddies mainly develop, exist and decay at the level where they formed in each case. There is thus almost no exchange of momentum between layers, so that the frictional effect of the sea or land is almost wholly exerted upon the air closest to the surface, with the result that the surface wind is considerably less than the geostrophic wind at, say, 2,000 ft.

Above 2,000 ft the variation of wind speed and direction is largely dependent on horizontal distribution of temperature. Within the region of 'Westerlies' of both hemispheres winds normally increase in strength up to the height of the tropopause. (*See* page 2.)

Pressure exerted by Wind

Figures for pressure exerted by the wind in lb/ft^2 at standard air density are included in the Beaufort Force tables. This wind pressure is roughly proportional to the square of the wind speed. The mechanical effect of wind on any structure, however, depends not only on the mean wind speed but also on the short-term variations of speed and direction due to gusts and squalls. A structure, such as a bridge, may be perfectly safe while supporting a steady wind pressure but unsafe when subjected to a series of blows, as from a gusty wind, particularly if the frequency of blows approximates to the natural period of vibration of the structure. Similar reasoning will apply in considering the risk of a ship 'dragging' in heavy weather in an exposed anchorage, or the more uncommon case of a ship listing due to shift of cargo or instability and experiencing a gusty wind on the beam, or in the case of a sailing vessel which is exposed to sudden gusts.

Trade Winds and Monsoons

The trade winds and monsoons illustrate the applications of Buys Ballot's Law, and the effect of the rotation of the earth on the movements of air on a large scale. The main wind flow extends up to about 10,000 ft. (*See* page 79.)

Local Winds

Various local winds, including land and sea breezes, katabatic and anabatic winds are described in Chapter 7.

Waves (Sea and Swell)*

The mariner lives in intimate contact with the waves of the sea and is able to realise better than most people the extent to which their size and energy, as shown by their destructive power, are related to the speed of the wind. He is also accustomed by his training to make frequent estimates of wind force and to use various terms by which to describe the state of the sea surface, although these may not be quite the same as the definitions accepted internationally for use by meteorologists and oceanographers.

Our knowledge of the way in which the wind produces waves on the surface of the sea is still incomplete. However, it is known that moving air exerts a drag upon a water surface with which it is in contact. Also, even in a light breeze, tiny ripples almost immediately form on a previously calm sea surface. Once wavelets have formed, provided the air flow is turbulent, as is normally the case, another process probably comes into operation. The pressure of the wind on the windward side of any wave will be greater than that on its leeward side, with the net result that the wave is driven forward in the direction of the wind. Thus, when a wind has been blowing in the same direction for some time, the waves of the sea are observed travelling in the same direction but with a speed somewhat less than the wind speed at the time.

A simple wave has four main properties. These are (*a*) SPEED, **C**, usually measured in knots, (*b*) LENGTH, **L**, measured in metres or feet, (*c*) PERIOD, **T**,

* *See* also *Marine Observer's Handbook*, Chapters 6 and 12.

expressed in seconds, and (*d*) HEIGHT, **H**, expressed in metres or feet. The speed refers to the rate at which individual waves travel; length is the horizontal distance between successive crests or troughs; period is the interval of time between the passage of successive crests past a given point; while the height refers to the vertical distance between the top of a crest and the bottom of a trough. In nature a train of simple waves on the sea (i.e. all the waves having the same speed, length, period and height) does not occur; instead the sea surface on most occasions presents a somewhat irregular appearance with waves of different height and length being found quite close together, although on many occasions the wave crests lie nearly parallel and move with almost the same velocity. There are two invariable relations which hold for simple water waves, when the water is deep, namely

$\mathbf{C} = 3{\cdot}1\ \mathbf{T}$ and $\mathbf{L} = 5{\cdot}1\ \mathbf{T}^2$ when the quantities are expressed in the units used above.

Experience has shown also that the largest wave heights usually occur with long wavelengths.

As a wind begins to blow over an area of some hundreds of square miles in extent where calm conditions prevailed beforehand, trains of waves are initiated which increase in height, period and length. While these waves are still in the same region they will continue to grow to a certain size which will be determined by the wind speed, the time for which the wind has been blowing, the distance over which the wind has blown (known as the FETCH) and the depth of the water. Initially the waves will be low and steep, that is their height/length ratio will be relatively large compared with a later time when, although they will have increased in height and length, the value of their height/length ratio will have become much less than it was earlier because their length will have increased to a proportionately much greater extent. The waves we have so far been considering are those directly created by the wind in the area; in other words, they are SEA waves. Their characteristics are intimately related to the force of the wind (see the modern specification of the Beaufort wind scale) and their appearance would be colloquially referred to (e.g. in the ship's deck log) as 'smooth sea', 'rough sea' or 'high sea', according to their height. However at some later time a stage occurs when the wind dies down over the area, and as a result the train of waves which it has created will soon travel away from that area. What happens when the wave train receives no more energy? In practice wave trains will travel for some time before their height becomes negligible and at the same time their wavelengths will increase slowly. These kind of wave trains which are encountered beyond the region in which they were generated are known as a SWELL and frequently travel thousands of miles before dying away. A swell wave in the open sea is distinguishable from a sea wave because the former has an oily, unbroken surface, whereas a sea wave invariably 'breaks'. Swell waves are referred to colloquially (e.g. in a ship's deck logbook) as 'low', 'moderate' or 'heavy', according to their height.

For meteorological purposes, both 'sea' and 'swell' are referred to as 'waves' but the direction, height and period of 'sea' and 'swell' waves are recorded separately (details are given in the *Marine Observer's Handbook*). There are obvious advantages in recording actual dimensions of the waves in this manner —even if these are only estimates, and the present system is that Voluntary Observing Ships record the direction, height and period of the waves they observe.

Propagation of Waves

At this point it is useful to return to the case of a water surface across which a train of simple waves is moving. During the period (T) of a wave each wave has moved into the position previously occupied by the wave before it,

$$\text{i.e. wave speed} = \frac{\text{length}}{\text{period}}.$$

In fact observation of a cork on the surface of the water shows that though the waves move, the surface of the water does not move with the waves. Instead the cork, and every other point on the water surface when waves are present, describes a circle in a vertical plane, moving upwards as the crest approaches, forward as the crest passes, downwards as the crest recedes and backwards as the trough passes. The diameter of this circle is equal to the height of the waves, while the time in which the cork moves once around it is equal to the period of the waves. The movement of every point on the surface of a water wave similarly takes place in a vertical circle, with the result that the profile of the wave surface is known as a 'trochoid', which is the curve that would be traced by a chalk mark made on the tyre of a bicycle as the wheel rolls along the ground. (The movements of water particles in waves are discussed more fully in the *Marine Observer's Handbook,* Chapter 6.)

The disturbance of the surface which constitutes a travelling wave extends below the surface. The water particles affected by the disturbance all move in circular orbits, whose diameters rapidly decrease with depth. Submarine crews know that the sea is virtually undisturbed at a depth equal to the length of the waves at the surface; in fact the diameter of the circles described by water particles at that depth is about 1/600 of the diameter of the circle described by a water particle on the surface at the same time.

Energy of Waves

A train of waves in the ocean has potential energy due to the elevation and depression of the surface from its initial state of being level, and also kinetic energy due to the movement of every particle in a vertical circular orbit described earlier. Theoretical reasoning shows that the amounts of potential and kinetic energy are equal and proportioned to $\mathbf{LH}^2$ per wavelength per unit of crest length. This means that waves 25 ft high will transfer energy to a coastline equivalent to about 230 h.p. per foot length of coast, which explains the destructive power of the sea upon cliffs, and upon man-made structures such as breakwaters and ships aground on sandbanks or rocks.

The energy of deep-water waves moves in their direction of travel with only half of the wave speed. This fact is responsible for the propagation of trains of waves with only half the velocity of the individual waves. In any group of waves the ones in the middle have the largest amplitude (i.e. wave height), while those at the front and rear have small amplitudes which decrease to nothing at the edges of the group. While the group is travelling the waves are continually overtaking the front of the wave train where they disappear, and thus the 'group velocity' of a particular group of waves is one-half the velocity of the individual waves.

Waves lose energy when they break at the crest with the formation of white horses, or when the wind begins to blow with a component from the opposite direction. Some of their energy is consumed in overcoming the internal friction (known as 'viscosity') which is evident between masses of water moving relative to each other. Treacle, for example, is a very viscous substance in which it is difficult to produce wave motion. The viscosity of water is small and this means that wave trains in water can travel great distances before they are finally dissipated.

Forecasting Wave Characteristics

The need often arises, and most urgently in time of war, for accurate forecasts of wave height. As we have already seen, the main factors which influence wave height are the velocity of the wind and the time for which it has been blowing. Often in stormy regions such as the North Atlantic in winter and the Southern Ocean at all seasons, the wind will rise from near calm to gale force or more within an hour or two. The time taken is often less than that required for the resultant sea waves to grow to the maximum height possible at such a wind speed. Tables have been worked out by oceanographers in various countries, from which wave height in the open ocean can be determined, when the wind speed and the time for which the wind has maintained this speed are known. All these tables assume that the wind speed and direction are constant during the period which is being considered.

When the need arises to forecast or estimate the probable wave height in a position somewhere near a coastline, the direction of the wind becomes important, if it is off-shore. This is because a wind needs to blow over a certain length of water known as the 'fetch' of the wind, before it is able to create waves of the maximum height possible for a wind of that velocity when blowing for an unlimited time. It follows that in making every forecast or estimate of wave height for a particular locality and occasion, account must be taken of the limitations or the theoretically possible wave height which are imposed by the *time* for which the particular wind has prevailed in the area, and the *fetch*, or distance of sea over which it has been able to act. Limits are also imposed on the wave height by the fact that a wave cannot exceed a steepness of $\mathbf{H}/\mathbf{L}>1/13$ without breaking at the crest, and similarly when the depth of water is much less than $\mathbf{L}/2$ breakers are formed.

The tables in use give the mean height of waves formed, which implies that they must be used with discretion, since statistics show that a small proportion of the waves may reach roughly twice the average height even in deep water.

The steepness $\mathbf{H}/\mathbf{L}$ of waves cannot exceed a value of about $1/13$, because up to this value of steepness any excess energy received from the wind is absorbed and increases their height relative to their length, but beyond it the excess wind energy is dissipated by the waves breaking at the crests with the formation of white horses. The highest waves can only occur with a sea of very long wave-lengths and these facts serve to explain why waves exceeding 50 ft are rarely observed.

Wave Observations and their Value

Although wave observations have been made by observers upon moving ships for many years past, and during the last few years from 'stationary' ocean weather ships, our knowledge of sea and swell waves is far from complete even on the most frequented shipping routes. The information relating to deep-water waves which has been collected by observers using methods of estimation described in detail in the *Marine Observer's Handbook* has, during recent years, been supplemented by observations with a wave recorder which has been installed in a British ocean weather ship. This recorder has the great advantage over all previous ones that it is contained entirely within the ship. The trace given by the instrument is a complete record of waves of all lengths which pass the ship during the period when the instrument is being operated. Short-period waves, which are of small amplitude, are recorded by means of the pressure exerted by the varying head of water at two small openings at the same level below the water-line, on opposite sides of the ship. Long-period waves are recorded by measuring the varying vertical accelerations to which the whole ship is subjected by these waves. These accelerations are taken from the varying periods of oscillation of pendulums in the instrument, which are fed into an electronic device which integrates the readings twice to obtain the true wave profile. In spite of practical difficulties connected with the pendulums and electronic equipment, the instrument has given valuable results which are in excellent agreement with observations made by other instrumental methods. The maximum wave height recorded so far (1965) in the North Atlantic is 67 feet by a wave recorder aboard a weather ship.

At the present time much remains to be learned regarding the production, travel and decay of ocean waves, information which is necessary before accurate forecasts of wave characteristics can be made available. Knowledge of height and other characteristics of waves is of considerable practical value for a variety

Table 3.2. Gradient wind speeds for various values of geostrophic wind and radius of curvature of isobars (Lat. 55°)

Geostrophic wind speed (knots)	Radius of curvature of isobars (nautical miles)											
	25	50	100	150	200	300	400	500	750	1000	1500	2000
	CYCLONIC CURVATURE Gradient wind speed (knots)											
5	4	4	5	5	5	5	5	5	5	5	5	5
10	6	7	8	9	9	9	9	10	10	10	10	10
15	8	10	12	13	13	14	14	14	14	15	15	15
20	10	13	15	16	17	18	18	18	19	19	19	20
30	13	17	20	22	24	25	26	27	28	28	29	29
40	16	20	25	28	30	32	33	34	36	37	38	38
50	18	24	30	33	35	39	40	42	44	45	47	47
60	21	27	34	38	41	45	47	49	52	53	55	56
70	23	30	37	42	46	50	53	56	59	61	64	65
80	24	32	41	46	50	56	59	62	66	69	72	74
90	26	35	44	50	55	61	65	68	73	76	80	82
	ANTICYCLONIC CURVATURE Gradient wind speed (knots)											
5				5	5	5	5	5	5	5	5	5
10				12	12	11	11	11	10	10	10	10
15				24	19	17	17	16	16	16	15	15
20					32	25	23	22	21	21	21	20
30						47	39	36	33	32	32	31
40							63	53	47	45	43	42
50								79	62	58	55	53
60									80	72	67	65
70									103	88	80	77
80									147	106	94	89
90										128	108	102
	25	50	100	150	200	300	400	500	750	1000	1500	2000

of purposes, including the design and behaviour of ships at sea; the design and orientation of harbours and the construction of breakwaters; problems of coast erosion and silting; discharging of ships in open anchorages; landing operations on exposed beaches; and the landing and take-off of civil and military aircraft, either with reference to the sea itself or to carriers. For example, the behaviour of an individual ship at sea is governed to a considerable extent by the period of her roll and pitch in various conditions of loading in relation to the period of the waves she encounters, and her longitudinal and transverse strength calculations must inevitably take account of similar factors. A modern need for accurate and frequent wave observations and for forecasts of waves (sea and swell) is in connection with the weather routeing of ships (*see* page 187).

In tropical waters the arrival of quite a gentle swell may be the first warning which is given of the approach of a dangerous tropical revolving storm. Similarly, wave data has value in areas where direct observation of storms are not available because of the limited amount of shipping in the area. From time to time public enquiries are held into shipping losses where the destruction of life and property was serious, and in such cases evidence provided by actual observations of wave height from ships in the vicinity is always of value.

CHAPTER 4

CLOUD AND PRECIPITATION

Cloud Form and Height

Clouds may assume an almost infinite variety of forms, but for description and identification these are classified as simply as possible. The present international cloud classification, which distinguishes 10 main forms, has been in use since 1896. Condensation of water vapour, which produces clouds, may occur in different ways, but the international classification is based on the appearance of the clouds.

Clouds occur almost wholly within the lowest layer of the atmosphere, the troposphere (page 2). They can only be formed in air which is cooled sufficiently to bring about condensation. Ascent in convection currents, upsliding at a frontal surface and, in certain circumstances, direct cooling of air by radiation are all processes which lead to condensation of water vapour. Whether condensation will occur or not also depends upon the air containing a supply of water vapour sufficient to result in the air becoming saturated before the fall of temperature is terminated. The presence of minute particles or nuclei in the air is also necessary for initiating formation of the water droplets or ice crystals which must precede the visible products of condensation, e.g. cloud, drizzle, rain and snow.

Clouds may be composed of water drops, ice crystals or a mixture of the two, according to the temperature of the air layer. High cloud is mainly composed of ice crystals, while medium clouds are largely formed of water drops. Low cloud usually consists entirely of water drops.

It can be shown from theoretical considerations that, in the absence of nuclei, water droplets or ice crystals cannot form from water in the atmosphere. Nuclei present in the atmosphere are produced by a number of processes but chiefly by the action of strong winds on salt spray which is blown away from breaking waves, by the products of industrial processes and less frequently by sand and dust carried to higher levels of the atmosphere by strong winds over desert regions and after volcanic eruptions. Recent researches indicate that large sea-salt nuclei are most important in promoting droplet formation, and that the lower levels of the atmosphere over the ocean contain many more sea-salt nuclei in conditions of strong wind and rough sea than when the wind is light and the sea calm.

On the polar plateaux of Antarctica and Greenland at air temperatures well below freezing point, ice crystals have been observed falling from cloud close to the surface. However, laboratory experiments have shown that the smallest droplets of pure water can be cooled to near $-40°C$ before freezing, and the presence of water-droplet clouds at these temperatures in the polar regions has been inferred with some certainty from observations of coronae and 'glories'. It is concluded that direct condensation to ice crystals does not occur until the temperature of the air approaches $-40°C$; at higher temperatures condensation to water droplets occurs on suitable nuclei which are always available in sufficient numbers in the atmosphere.

In the stratosphere the amount of water vapour is very small. Cirrostratus clouds occasionally form in the lower levels of the stratosphere, but cloud is seldom found at higher levels, apart from the rare 'mother-of-pearl' clouds which are described in the *Marine Observer's Handbook.*

Cloud is formed at levels ranging from close to the ground to a height of about 45,000 ft in temperate latitudes. As the troposphere extends to a greater height in the tropics, the highest cloud there may reach 50,000 ft or more. The highest clouds tend to be somewhat higher in all latitudes in summer than in winter as the tropopause is higher in summer than in winter at all latitudes, and consequently convection can raise water vapour to greater heights in the summer half of the year.

Since in a given locality condensation will occur at the same level, whether the air be rising in convection currents or by means of turbulence, it follows that the bases of individual cumulus, stratus or stratocumulus clouds will all lie at very nearly the same level. This arrangement of clouds in which the cloud bases of a particular type all lie at one height is well known to observers on sea or land, and can be verified more readily from a hillside or from an aircraft. However, the thickness of the various cloud types and the height of their tops are much more variable, particularly in the case of cumulus and cumulonimbus.

For simplicity, clouds are grouped broadly according to their heights into high (or upper), medium (or middle) and low clouds. Clouds of great vertical extension have their bases at 'low cloud' height and their tops at 'medium' or 'high cloud' height.

The following table shows the main cloud forms and the average range of heights of high, medium and low types of clouds.

Table 4.1. Cloud forms and heights (temperate latitudes)

High Clouds	Medium Clouds	Low Clouds	Low Clouds of Marked Vertical Extent
45,000 ft to 18,000 ft	18,000 ft to 6,500 ft	6,500 ft to close to ground	45,000 ft to close to ground
Cirrus (Ci) Cirrocumulus (Cc) Cirrostratus (Cs)	Altocumulus (Ac) Altostratus (As)	Nimbostratus (Ns) Stratus (St) Stratocumulus (Sc)	Cumulus (Cu) Cumulonimbus (Cb)

The average height at which each cloud type may form can thus vary considerably. Furthermore, the limits given for high, middle and low clouds are not precise, e.g. medium clouds may occur a little below 6,500 ft and high clouds below 18,000 ft.

Cloud Classification

The original classification, made by Luke Howard in 1803, contained four types, namely cirrus, cumulus, stratus and nimbus. Cirrus refers to a silvery fibrous or feathery high cloud, seen in blue sky. Cumulus is a white, cauliflower-like cloud, sometimes of considerable vertical height. Stratus is a more or less even layer of grey cloud, not giving rain, while nimbus, which is now termed nimbostratus, is a darker layer, giving rain. Although not detailed enough for modern requirements, this classification is the basis of the one now in use. Thus the name 'cirrocumulus' is formed from 'cirrus and cumulus'. These combined names refer to clouds intermediate in form between two of Luke Howard's types. Exceptions are altocumulus and altostratus, in which the prefix 'alto' is used to denote 'of medium height'. Luke Howard's original nimbus is now termed nimbostratus.

Most clouds can be placed into one of two classes:

(*a*) Clouds in the form of layers or sheets.

(*b*) Clouds of considerable vertical thickness, which is comparable with their horizontal extent (heap or cumuliform cloud).

This distinction is related to the way in which the cloud is formed and helps in learning the main cloud forms. A cloud in layer form may cover the whole visible sky or only a part of it; the flatness of the cloud in this form is usually obvious, even when it is of small extent. The main point to note is whether the layer of cloud, whatever its extent, is wholly or mainly unbroken, showing structureless uniformity; or is composed of more or less regular and rounded individual cloudlets, separated from one another and between which blue sky may be seen. If of the former appearance, it is a stratus type and this word will occur in its name; otherwise it is of a cumulus type and that word will occur in its name.

The chief distinctive features of the main cloud forms are given below:*

1. Cirrus (Ci). Detached clouds of delicate fibrous or feather aspect in blue sky. They are generally white, often of a silky appearance, and exhibit no shading. Cirrus appears in the most varied forms, such as isolated tufts, branching plumes, straight lines, curved lines ending in tufts, etc. They are often arranged in parallel bands, alone or mixed with cirrostratus, which appear to converge to the horizon, or to two opposite points on the horizon, by perspective effect.

Cirrus appears as scattered thin cloud in blue sky, and the sun shines through it with nearly undiminished intensity. While the feathers or patterns of its individual cloudlets may in some cases unite in a curved or branching pattern, it cannot cover the sky uniformly either as a continuous layer or as one broken up into cloudlets of about the same size. It may, however, gradually change into either cirrostratus or cirrocumulus.

'At night the brighter stars can be observed through cirrus and cirrostratus, the presence of this cloud causing a slight diffusion of the light of the stars shining through it.

2. Cirrocumulus (Cc). A layer or patch of cloud of cirrus character, composed of very small separate rounded cloudlets or small white flakes, without shadows. These are arranged in groups or lines, or more often in ripples like those of sand on the seashore. (*See* Altocumulus.)

3. Cirrostratus (Cs). A thin whitish veil, which is sometimes so diffuse as merely to give a milky appearance to the blue sky. At other times it may show a fibrous structure, in greater or less degree, with disordered filaments, but is never thick enough to blur the outlines of sun or moon. It gives rise to halo phenomena. Cirrostratus, alone or mixed with cirrus, is often arranged in parallel bands which appear to converge to the horizon by perspective effect.

* Photographs of these cloud forms can be found in the *Marine Observer's Handbook*.

4. Altocumulus (Ac). A layer or patch composed of cloudlets, each of which has a flat or flattened globular appearance. Blue sky usually shows between the cloudlets, but these may be so close together that their edges join. They are arranged in a more or less regular pattern of groups, lines or waves, aligned in one or two different directions. The separate cloudlets may or may not show shadow; their edges are thin and semi-transparent. Their size varies considerably, in general it is intermediate between that of cirrocumulus and stratocumulus cloudlets. Altocumulus, at moderate angular distances from the sun, may show iridescent colouring—delicate tints of red and green and sometimes other colours of the spectrum. The name 'mackerel sky' applies to cirrocumulus or altocumulus formed of comparatively small cloudlets. (*See* Stratocumulus.)

5. Altostratus (As). A striated or fibrous veil, which is grey or bluish in colour and has a resemblance to a veil of cirrostratus, although generally it is rather thicker. It does not give rise to halo phenomena. In its thinner forms, it blurs the outlines of the sun or moon, as if they were viewed through ground glass, giving the appearance known as a watery sky. In its thicker forms, the sun or moon shows only as a vague blur or it may be completely hidden. Thick altostratus usually shows relatively light patches between darker ones but the surface never shows real relief of light and shade, as altocumulus does.

6. Stratocumulus (Sc). A layer or patches composed of globular masses or elements, which appear soft and grey, with relatively dark shadows. These elements are arranged in a more or less regular pattern of groups, lines or waves, usually aligned in one direction but occasionally in two different directions. The elements are generally closer together than those of altocumulus and their edges often join together, but small streaks or patches of blue sky may be seen between them. Frequently they are completely joined to form an overcast sky, distinguished from stratus by having a wavy or linear appearance. A special form of stratocumulus, long parallel rolls of clouds, showing blue sky or higher clouds in the gaps, is called 'roll cumulus'. The usual height of stratocumulus base is between 1,500 ft and 4,500 ft.

7. Stratus (St). A more or less continuous sheet or layer of cloud of low height, resembling fog, but not resting on the ground. When broken up by wind, or otherwise, into patches or shreds, it is called stratus fractus or cumulus fractus. The usual height of the base of stratus is between 500 ft and 2,000 ft.

8. Nimbostratus (Ns). A low rainy layer, of nearly uniform structure and of a dark grey colour, as if faintly 'illuminated from inside'. It may or may not give rain or snow; if it does, precipitation is continuous. Its lower surface may have a diffuse or 'wet' appearance, owing to straight or curved lines of precipitation trailing from it. Such precipitation may or may not reach the ground. The darker and often threatening appearance, the 'wet' look and also the precipitation, if this occurs, distinguish this cloud from stratus. The base of nimbostratus is usually between 500 ft and 2,000 ft but it may come down nearly to the earth's surface. Sometimes it is as high as 4,000 ft.

The special feature of nimbostratus is that, while it appears as a layer from below, its thickness cannot be gauged because of its opacity. The fact that it obstructs more light than other layer clouds shows that is has considerable thickness and is not a true layer cloud. It may be the under-surface of a cumulonimbus, thousands of feet thick, or it may extend vertically to meet altostratus above it, thus forming a very thick composite cloud which constitutes the central cloud core of a depression.

Ragged low clouds of bad weather, called stratus fractus or cumulus fractus may often be seen below nimbostratus, or below altostratus. These are called 'scud' by seamen, because their movement in the strong winds of a depression often appears very rapid, due to their low height.

9. Cumulus (Cu). A cloud with a base that is horizontal, or nearly so, and a rounded or dome-shaped top, usually with protuberances. Strong contrasts of light and shade may occur on these clouds; against the sun they appear dark, with brightly luminous edges. True cumulus is definitely outlined, often appearing very hard and clear-cut and brilliantly white against the blue sky. Cumulus clouds vary greatly in general size and vertical height, and all degrees of size and massiveness may be observed from a small cumulus up to the great cumulonimbus described below. The usual height of the base of the cloud is from 1,500 ft to 5,000 ft.

Cumulus clouds of no great thickness and having ragged edges are called cumulus fractus. This form can usually be distinguished from stratus fractus by having a less flattened appearance.

10. Cumulonimbus (Cb). A heavy mass of cloud of great vertical height, whose summits have the appearance of rounded mountains or towers, appearing very bright by reflection of sunlight. In its most developed form, the upper part of the cloud spreads out in the shape of an anvil of brilliant white cloud, with a nearly horizontal upper surface and a fringe of fibrous cloud all round, resembling cirrus. This fringe is known as anvil cirrus (formerly called 'false cirrus'). The base of cumulonimbus resembles nimbostratus, and generally gives showers of rain or snow and sometimes hail or soft hail. Thunderstorms of varying degrees of severity may occur. If the shower cloud cannot be distinguished, the occurrence of a shower, as opposed to continuous rain, will characterise the cloud from which it falls as cumulonimbus. The usual height of the base is the same as that of cumulus; the top of the cloud may extend to cirrus level.

Large cumulus is sometimes called 'woolpack' or 'cauliflower' cloud; cumulonimbus is known as the thundercloud or shower-cloud.

The effect upon daylight of the four 'stratus' cloud forms varies as follows:

(*a*) Cirrostratus. Sun only dimmed a little and still throws shadows.

(*b*) Altostratus. Shows the sun as a watery blur which does not throw shadows. In thicker forms of the cloud the sun may become almost or quite invisible.

(*c*) Stratus. Sun wholly obscured. Daylight weakened and diffused.

(*d*) Nimbostratus. Dark and threatening. Daylight greatly reduced.

Special Varieties of Cloud Form

The international classification of cloud form includes a number of sub-types and varieties of each main form. Most of these are not mentioned here as the marine observer has no need to know them. A few, however, are well known and distinctive and are given below.

Anvil Cirrus. In the movement or dissipation of cumulonimbus the 'anvil' of cirrus may become detached. Such anvils may be seen in blue sky after thundery weather, with no cumulonimbus clouds in sight. They will probably have lost much of the anvil shape and are distinguished from ordinary cirrus in forming denser patches; in this respect they resemble patches of cirrostratus but have a softer outline.

Lenticular cloud. Clouds sometimes assume hard-edged oval or lens-shaped forms popularly known as 'whaleback' clouds. This may happen with different cloud types at all heights, but is most common with altocumulus. (Altocumulus lenticularis (Ac lent.).) Such clouds are generally associated with a bright blue sky, a strong gusty wind and an unsteady barometer. Usually only a small number are seen at one time, often widely spaced.

Altocumulus castellatus (Ac cast.). This is a type of altocumulus in which individual cloudlets extend vertically in turrets or castles. It is not distinguishable overhead; it is best seen in profile at moderate or low elevation. It often precedes thundery conditions.

Mammatus cloud. This cloud form, when well developed, gives a most unusual appearance to the sky. The distinguishing feature is that the lower surface of a cloud of mammatus type hangs down in udder-like bulges of soft outline but showing relief of light and shade. These bulges are caused by down currents of air. It may occur with cloud of various types and heights but chiefly below the base of a cumulonimbus or cirrus anvil, when it is called mammato-cumulus. It thus occurs mainly in thundery conditions.

Formation of Cloud

The formation of cloud results from condensation of water vapour in the free atmosphere; when such condensation happens near the surface mist or fog is produced, which can thicken upwards to a depth of several hundreds of feet. Sometimes the base of low cloud falls to the surface, commonly over high ground or at a snow-covered surface, where it reduces visibility at ground level in the same way as fog does. Cloud, on the other hand, forms at various heights in the troposphere (up to about 45,000 ft). This is the only distinction between fog and any cloud composed of water drops. Cloud is generally higher above the earth's surface than fog, but not invariably; the base of very low cloud may be on the ground while the upper surface of a high stratum of fog may sometimes be as much as 2,000 ft to 3,000 ft above the ground. The formation of fog will be discussed in Chapter 5.

As explained on page 50, condensation to water drops or ice crystals is caused by the cooling of air to a temperature below the dew-point or frost-point. Air may cool in various ways, but practically all cloud formation, as distinct from that of fog or mist, occurs as a result of adiabatic cooling. This occurs, as explained on page 12, when the pressure within a parcel of air is reduced by ascent which can be either spontaneous or forced. Cloud formation, therefore, results from the rising of moist air. There are four ways in which such a rise may occur.

(*a*) By turbulence in strong winds, especially when blowing over a cooler surface. In such winds the vigorous stirring causes a more uniform distribution of water vapour in the surface layers and, if the surface is a cold one, a general cooling of those layers, with the result that condensation may occur to form cloud at a fairly low level.

(*b*) By orographic ascent (Greek *oros*, mountain), caused by obstruction to the wind on a larger scale than that described as turbulence. When the wind blows against hills or mountains some of it is forced to ascend over the mountains.

(*c*) By convection. When a portion, or parcel, of air becomes locally heated to a higher temperature than the surrounding atmosphere, it will rise freely through the atmosphere if the lapse rate exceeds the dry adiabatic up to the condensation level.

(*d*) By general ascent or upsliding. It will be shown in Chapter 9 that the following air movements occur during the passage of a depression. Forward of the depression a warm air mass approaches a colder air mass at a warm front, and the warm air rises up over the cold air all along the surface of separation. In the rear of the depression, a cold air mass meets a mass of warmer air and undercuts it at a cold front, forcing the warm air upwards. In these conditions, ascent of air can occur over considerable areas and be associated with rather extensive areas of precipitation. The ascents of air produced under the other headings (*a*) to (*c*) are normally on a smaller scale.

These processes will now be considered further.

TURBULENCE CLOUD. In the lowest layers of the atmosphere turbulent eddy-motion produces a uniform distribution of water vapour, with the result that the dew-point tends to be uniform throughout these layers. If the humidity is high enough, the temperature of the highest part of the mixed layer will be below the dew-point, while that of the air near the ground is above the dew-point. Cloud will form at the height at which the dew-point is reached. Turbulence cloud may take the form of stratus fractus or may be thick stratus or stratocumulus in a continuous layer. Turbulent mixing cannot extend to any great height, so the base of the cloud is low, generally below 2,000 ft. The process is illustrated in Fig. 4.1.

Over the land the most rapid variations in relative humidity and turbulence occur in the early evening and again after dawn, and it is at these times that changes in turbulence cloud are most likely to occur, either new cloud being formed or previous cloud clearing. Over the open sea little diurnal variation occurs in cloud types and amounts.

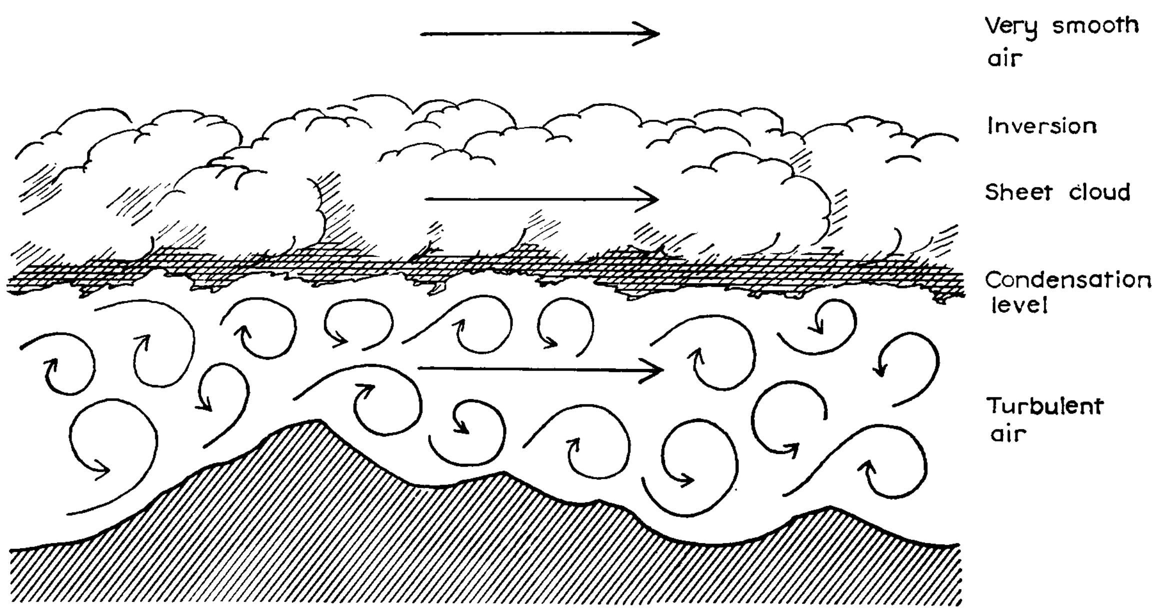


Fig. 4.1. Formation of a Layer of Turbulence Cloud

Turbulence cloud is common when air of subtropical oceanic origin reaches temperate latitudes. Such air is initially rather moist and on reaching the cooler region its relative humidity becomes still higher. In these circumstances layers of low cloud can be persistent over the North Atlantic at any season, and also over the British Isles, although inland in summer they often become well broken or disappear altogether during the day.

Stratus fractus or cumulus fractus (scud) of bad weather is turbulence cloud formed below the main cloud mass of a depression as described earlier on page 54.

OROGRAPHIC CLOUD. This may be observed at sea over high islands or over mountains near the coast. The 'tablecloth' of Table Mountain and the Levanter cloud over Gibraltar are orographic clouds.

Orographic cloud is often of stratus type with a flat base and only occurs if the air cools enough in ascending to bring its temperature below the dew point. When formed on a small scale, for example over an isolated ridge or hill or over a single small hilly island, it has a characteristic 'laminated' shape resembling a flat plate, and sometimes it can be seen that the cloud is maintaining its position while the air is passing through it. In effect the cloud is continually renewed by air entering it on the windward side and is at the same time continually dissolved by air leaving it on the leeward side, as shown in Fig. 4.2 The island or hill then has a cap of cloud which may cover the summit or lie above it but the cloud does not move since, as already explained, the wind blows through it.

Warm moist air blowing against a high mountain range forms orographic cloud on a large scale, which often develops into nimbostratus and is accompanied by heavy and prolonged precipitation.

CONVECTION CLOUD. From its origin this cloud assumes one of the forms of cumulus with a large or small vertical extent. The convection currents may be caused by local heating of air in contact with the ground, as during a warm summer day on land; in the trade wind region, the warm sea causes convection

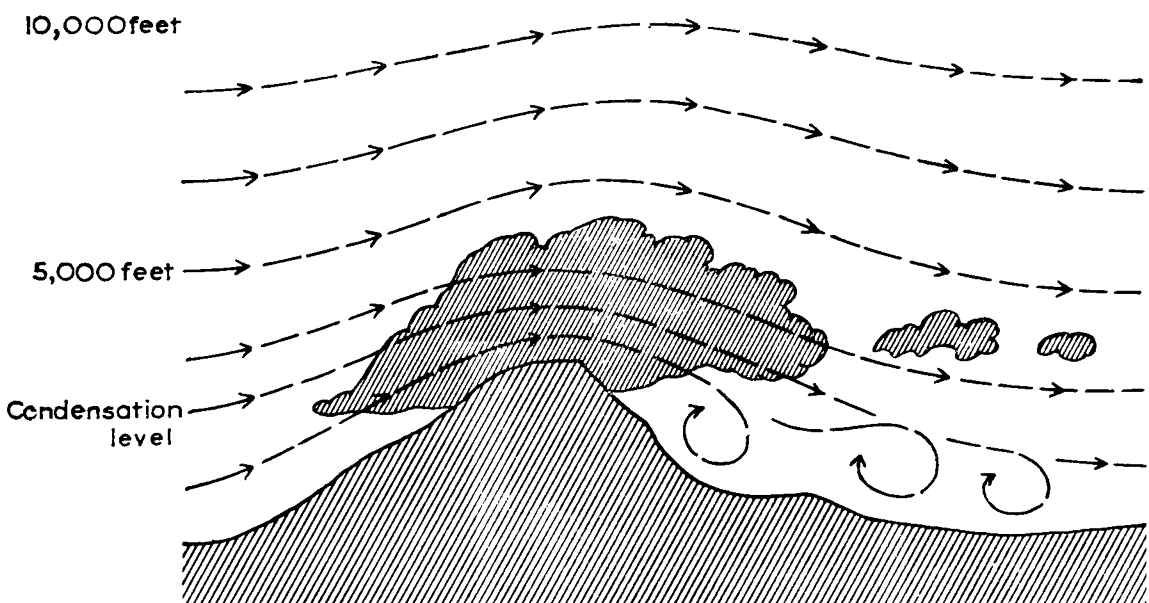


Fig. 4.2. Formation of Orographic Cloud

currents in the lower layers of the cool air coming from higher latitudes. The whole of this warmed air ashore or at sea cannot rise as a single mass over a large area, as other air must fall to take its place. The air thus ascends in separate columns and, wherever a column has penetrated the condensation level, the space occupied by the rising air above that level is filled with cumulus cloud. This rising in isolated columns is facilitated ashore by the unequal heating of the ground, dependent on the nature of the surface and its vegetation covering.

A small island often causes ascending columns of air to develop over it on a sunny day and become capped with cumulus cloud as the day advances. If there is a moderate breeze a succession of cumulus clouds will develop and drift many miles down wind before dispersing. In either case, whether the day is calm or windy, the typical convection clouds, which are a variety of cumulus cloud, are readily distinguished from the flat, lens-shaped orographic clouds which remain in the same position in all conditions of wind, as described earlier.

Convection cloud may also form along an extensive coastline with an onshore wind, even when the air further to seaward is cloudless. This cloud formation is caused mainly by the air being heated to a greater extent over the land than over the sea, and to some extent by uplifting of the air if there are cliffs or hills near the coastline.

In the trade-wind regions and on summer days ashore in temperate latitudes, cumulus clouds increase in thickness only slowly if at all, unless the atmosphere is markedly unstable. In the doldrums, on the other hand, once cumulus have been formed they often extend rapidly upwards.

When convection is vigorous and occurs through a deep layer, a cumulonimbus may result, its head towering into the high cloud levels. When the air at the top of the column cannot rise any higher, its continued ascent results in a horizontal spreading out of air from the top of the column. The water vapour in this spreading layer forms the ice-crystals of the 'cirrus anvil', whose flat top marks the limit beyond which, in the prevailing conditions, further rise of air is impossible.

The size and type of cumulus cloud are so varied that it is useful to divide them into five classes.

(*a*) Small cumulus, relatively flat in appearance, often with irregular or serrated edges. The dome shape is not well developed. This is the normal cumulus of fine summer weather in temperate latitudes. Height of base 1,500 ft to 5,000 ft.

(*b*) Cumulus with rounded top, extending up to several thousand feet, the general appearance being dome-shaped. Height of base 1,500 ft to 5,000 ft.

(*c*) Larger cumulus of greater height, often with multiple domes, which have a cauliflower appearance as if they were swelling and sprouting upwards. The base still remains flat, with no sign of rain falling from it.

(*d*) Cumulonimbus, of massive form and great height. The general shape varies very much and is often irregular. Cumulonimbus without anvil cirrus is distinguished from cumulus of class (*c*) by having a dark base of nimbostratus from which rain may fall, or be seen to trail downwards without reaching the ground.

(*e*) The final development of cumulonimbus with a cirrus anvil forming or formed.

The larger cumulus and the cumulonimbus, classes (*c*) to (*e*), have the base at about 2,000 ft; however, the base of cumulonimbus is frequently lowered due to the underlying air becoming saturated by rain falling through it. The cloud thickness in class (*c*) may be 10,000 ft or more, while cumulonimbus clouds extend in height to 6 or even 9 miles in the tropics.

Over the land fine-weather cumulus clouds are subject to a marked diurnal variation. They usually form in the morning and reach a maximum in number and size in the afternoon. They dissipate rapidly when the ground cools in the evening and convection currents die out. Over the sea diurnal variation of temperature is much less marked, and any variation in cloud tends to be the reverse of that over land. The sea temperature remains constant, while the air aloft tends to cool at night by radiation. Thus convection and cloud formation may increase a little by night and decrease a little by day.

Cumulus fractus does not appear in the above summary of cumulus types since it is not a separate type of cumulus, but rather cumulus cloud of classes (*a*) and (*b*) which is being broken into smaller portions by strong wind or in process of disintegration due to evaporation. Cumulus fractus has a characteristic ragged appearance; sometimes it is indistinguishable from stratocumulus disintegrating under the same influences.

CLOUD FORMED BY GENERAL ASCENT OR UPSLIDING OF AIR OVER A WIDE AREA. Formation of cloud in this way occurs mainly in the frontal regions of a depression. The ascent is mainly caused by a warm air mass flowing up a gently sloping frontal boundary separating it from colder air. The air either ascends gradually, as in the case of a warm front, or more violently in the case of a cold front where the colder air undercuts the warm and the slope of the frontal boundary is steeper than at the warm front.

Cirrus, cirrostratus, altostratus and nimbostratus are thus formed at a warm front; and cirrocumulus, altocumulus, cumulus and cumulonimbus at a cold front.

General ascent of air, due to convection on a large scale, occurs with the formation of polar and tropical depressions which do not possess fronts. Many extensive belts of more or less continuous cloud and rain are produced in this way.

Cloud Movement and Changes

CLOUD MOVEMENT. Clouds move generally with the speed and direction of the wind at the height where they are situated; thus, they may move in a direction and speed other than that of the surface wind. Sometimes when there are clouds at different heights they can be seen to be moving in different directions. To an observer the apparent rate of movement of clouds at a low level across the sky is often larger than that of higher clouds, because the low cloud is so much nearer to the observer.

In areas of light winds, the movement of cumulonimbus of the thunderstorm type depends more on individual cloud development than on wind drift. The movement of a group of such clouds as a whole may be erratic and unrelated to the surface wind. They may tend to spread outwards and degenerate, or break up into two or more separate storms.

CLOUD CHANGES. The clouds are continually changing in form or detail or both. The changes of a cumulonimbus are rapid and obvious, but it is difficult to see changes in structureless, overcast skies. Apparent changes result from variations in the distance and angle of view, and in lighting due to the sun or moon, and may also result from the cloud's movement.

Real changes are due to:

(*a*) Continuing growth by condensation, without change of cloud type.
(*b*) Gradual clearance of the cloud by evaporation, without change of type.
(*c*) Change of cloud from one type to another, e.g. from cirrus to cirrocumulus or from altocumulus to altostratus. Such changes may or may not be accompanied by general growth or decay.
(*d*) Deformation by wind action.

A commonly observed sequence of cloud change is from cirrus to cirrostratus, followed by altostratus and finally nimbostratus as a warm front approaches the observer. But the clouds in this case are not strictly changing, they are part of a large system of cloud which is moving towards and past the observer.

Some clouds change to other cloud forms at the same general height, while others change both in type and height. Thus, when altostratus changes to nimbostratus the cloud base gradually lowers from the height of medium to the height of low cloud due to condensation at lower levels; when nimbostratus changes to altostratus, the base gradually rises due to clearing of the cloud at lower levels.

Precipitation

The term PRECIPITATION covers any appreciable fall or deposit of water drops to the ground, including its frozen or partly frozen states, which results from condensation of water vapour in the atmosphere. Mist, fog and cloud are forms of condensation but are not classed as precipitation, with the exception of wet fog, which deposits water on objects with which it has contact. The finest drizzle is precipitation and so is dew and the various forms of frozen deposit, hoar frost, rime, etc.

Formation of Precipitation—Rain

The water droplets or ice crystals whose formation results in cloud or fog, as discussed earlier in this chapter, are very minute in comparison with the drops composing drizzle or light rain. Before appreciable rainfall can result condensation must be assisted by other physical processes. Until a few years ago it was thought that, after a cloud had formed, the water droplets could only increase to the size of raindrops when in the presence of ice crystals. The argument was as follows. Since the saturation vapour pressure is lower over ice than over water, in a mixture of water droplets and ice crystals the latter would grow at the expense of the water droplets, falling towards the ground when they had grown large enough to attain the necessary fall velocity and melting to rain at a lower level where air temperature was above freezing-point. Precipitation in temperate latitudes is adequately explained by this process, since the vertical extent of most shower clouds is sufficient to allow upward currents of air in them to carry water droplets to a level where their temperature falls below $-10°F$, which is sufficient to promote formation of ice crystals and initiate the sequence of events just described.

In tropical regions heavy showers are quite common from clouds, the tops of which are known to be in air at a higher temperature than $-10°F$. In fact numerous observations have been recently confirmed in which showers occurred from clouds which were warmer than freezing-point throughout. It is now believed that this depends on the existence, at an early state in the growth of the cloud, of a few water droplets which are much larger than the majority of droplets of which the cloud is composed. These drops are able to grow rapidly by collision and absorption of the smaller cloud droplets; a crude analogy may be found in the marine world when a whale makes its way through a shoal of krill. In the case of the cloud droplets the process is self-sustaining in the conditions presupposed here; larger droplets falling faster than smaller droplets and absorbing many more of the very small drops in the process. Powerful up-currents further assist the process by lifting all but the largest drops and increasing the number of collisions. It is considered that these larger drops which are present at an early stage are formed on a small number of nuclei which are very much larger than the average size of the nuclei present.

These theories explain the main observed features regarding the formation of precipitation, which are that thin layers of cloud seldom give appreciable precipitation except in the form of drizzle, in which the water-drops are very small, and that light rain or snow seldom falls from cloud less than 3,000 ft thick. Clouds can be 10,000 ft thick without precipitation falling from them, and are often observed to have a vertical thickness greater than this before heavy rain starts to fall.

Although rain in showers is often heavy, the duration of showers is mostly limited to a few minutes. The raindrops found in showers are of all sizes up to the limiting size beyond which they cannot exist in a stable shape and must break up into smaller drops naturally. Steady precipitation is less intense but more prolonged and is usually formed from medium cloud forming part of the cloud system of a front.

Snow

Once a cloud droplet has frozen, further condensation is directly to ice. The ice crystal which is formed initially is very small and hence falls very slowly, so that there is plenty of time for the growth of the feathery crystalline ice structure which we call a snowflake to build up in its fall towards the earth. This structure examined under the microscope is seen to be of great beauty and infinite variety. Dry snowflakes do not rapidly combine, which explains why in temperatures well below freezing-point individual snowflakes are very small, and freshly fallen snow is easily lifted by the wind to be blown along in a fine cloud as in a blizzard. Large snowflakes are aggregates of crystals, probably resulting from the coalescence of crystals with water droplets. When the air and the ground temperatures are below freezing, snow remains unmelted as it falls. If the air near the ground is above the freezing-point the snow melts easily because the flakes fall so slowly, hence snow does not occur at temperatures more than a few degrees above the freezing-point.

Sleet is a mixture of snow and rain, or of partially melted snow, which occurs when air temperatures are within a narrow range a few degrees above freezing.

Hail

The formation of hail can only be briefly described. For a hailstone to grow to a large size very strong upward currents must develop, otherwise it will fall to the ground. These upcurrents can only occur when the atmosphere is very unstable, usually in cumulonimbus clouds, which is the reason why hail is often associated with thunderstorms. If a hailstone is split open it shows an inner soft core with a layer of clear ice outside; larger stones have several such layers. The first condensation in the upper part of the cloud is to supercooled water-drops. Later this freezes to ice crystals, which grow into soft white pellets by a process of condensation direct from vapour to solid. A pellet begins to fall as soon as it is heavy enough to overcome the resistance of the ascending air current. During its fall an ice coating is given by the freezing of the super-cooled raindrops the pellet encounters. The growing hailstone may then be carried up again by a strong air current and the process is repeated. This may occur several times, giving the several layers sometimes found. The hailstone finally falls to the ground when it becomes too heavy to be supported by the rising currents.

Thunderstorms*

When the air is very unstable and vertical convection has developed to such an extent that towering cumulonimbus clouds have formed, thunderstorms may occur. The mechanism of the thunderstorm is complex, and only a brief explanation is given below.

Intense convection leads to localised charges of electricity within a cumulo-nimbus cloud; enormous potential differences are set up, and discharges of electricity subsequently occur between two clouds or between a cloud and the earth. A theory suggested by Sir George Simpson to explain the accumulation of charges of electricity in the cloud maintains that when the raindrops supported in the cloud have grown beyond a critical size of about 0·25 in., they are torn apart by the air streaming past them. The breaking of the drops induces a positive charge on the small droplets so produced, and a negative charge in the surrounding air which is carried away to higher levels of the cloud in the rising air current. For the drops to reach the critical size, an upward current of about 1,600 ft/min is required, which is one reason why thunderstorms only occur when convection is extremely vigorous.

Thunder is simply the explosive report accompanying the flash of lightning and is caused by the sudden expansion of the air which has been made white hot. Since the flash and noise occur simultaneously, the distance of the storm from the observer may be estimated from the interval t, in seconds, elapsing between the time the flash is seen and the time the thunder is heard. Light travels at 186,000 miles/sec, which may be considered instantaneous. Sound travels about 1 mile in 5 seconds. Then $\frac{t}{5} = d$, where d is the distance of the storm, in miles, from the observer.

The long rumbling is often due to echoing of the thunder-clap from hills or other clouds. It may also be due to the fact that the thunder-clap is travelling different distances to the observer from a lightning-flash that is about a mile or more in length.

Summary of Classes of Cloud and Precipitation

(*a*) Turbulence cloud. Formed by discontinuous mixing between layers of air, near the surface, due to irregularities in the earth's surface.

* *See* also *Marine Observer's Handbook*, Chapter 11.

Occasionally a similar mixing occurs in a higher air layer, due to other causes. The cloud is low, with base generally below 2,000 ft and rarely more than 4,000 ft. It is of a layer type, stratus or stratocumulus or stratus fractus, and sometimes as much as 3,000 ft thick. Usually there is no precipitation; if it does occur it is very slight.

(*b*) Convection Cloud.

(i) When there is little instability at low levels, the prevailing type is isolated fair-weather cumulus cloud. The base is usually above 1,500 ft and below 5,000 ft. The cloud may be only a few wisps of cumulus fractus in some cases, or cumulus with a vertical thickness of several thousand feet, sometimes as much as 10,000 ft. Usually there is no precipitation; sometimes very slight showers occur.

(ii) When there is instability in the upper air, cloud development takes the form of large cumulus or cumulonimbus with or without anvil. The base is usually found at about 2,000 ft but may fall lower; vertical thickness of cloud is from 10,000 ft to 30,000 ft or more. Showers of heavy rain fall locally, perhaps with hail and thunderstorms. During decay of a thunderstorm patches or layers of clouds at all levels may form.

(*c*) Cloud due to a general rise or upsliding of air over large areas, during the passage of a depression. Clouds of all levels, from cirrus downwards, occur at some stage. More or less heavy rainfalls, sometimes with thunderstorms.

(*d*) Orographic cloud is developed by forced ascent of air by hills or mountains. It is of stratus type and gives little precipitation except in the case of high mountains.

For information about the effect of clouds and precipitation upon radar observations, *see* Chapter 5.

CHAPTER 5

VISIBILITY

General Remarks

Visibility describes the transparency of the atmosphere and may be defined as the maximum distance at which an object can be seen and its details discerned clearly enough for it to be recognisable in normal daylight.

Visibility depends chiefly on the number of solid or liquid particles held in suspension in the air. It may differ in different directions either because the concentration of particles varies between different directions, as in the neighbourhood of a large port or other industrial centre, or because of differences in lighting. For instance, visibility is usually better when looking away from the sun than when looking towards it, and better in strong light than in weak.

The main causes of atmospheric obscurity are:

(*a*) Visible moisture in the atmosphere. Under this heading are cloud, mist or fog consisting of water droplets, precipitation (i.e. drizzle, rain, sleet or snow, etc.) at the observer's level and spray blown up from the sea. Water vapour is a transparent gas and so does not affect visibility.

(*b*) Solid particles such as those produced by factories, domestic fires and forest fires, by sea spray and by sand and dust due to strong winds in desert regions, or as the results of volcanic eruptions.

Where particles from these sources are absent the atmosphere is nearly transparent, although the smaller nuclei when present in large quantities in a layer not necessarily near the surface cause a scattering of the light. This accounts for the pale blue or white skies sometimes seen in waters adjacent to desert areas.

Although there are several agencies whereby visibility in the atmosphere is reduced, over most of the world the commonest factor producing low visibility is the presence of water droplets. When atmospheric visibility is reduced below 1,100 yd due to water droplets, fog is said to occur. It frequently happens during duststorms, sandstorms and in industrial towns in winter time, that the suspended matter in the air reduces the visibility below 1,100 yd. Although, strictly speaking, these conditions are not aptly described as 'fog'; for making statistical summaries of visibility at a particular station all such occasions when visibility was below 1,100 yd are returned as fog or thick dust haze. In a synoptic weather report the visibility is indicated in metres (or yards) irrespective of the cause. Thus a reduction of the visibility to 200 metres may be caused by fog, snow or heavy rain; the cause will be indicated in the weather report under the heading of present weather.

Formation of Fog

A fog composed of water droplets may also be described as a cloud on the surface. Over high ground, fog may be merely one of the cloud types formed because of cooling by adiabatic ascent. In other cases, the condensation in a fog is almost entirely produced by the direct effect of a relatively cold surface. Two main types may be distinguished:

(*a*) RADIATION FOG, due to cooling of the ground and the air in contact with it, by radiation. It forms almost entirely at night and only over land, since the sea surface retains a fairly constant temperature. Radiation fog is, however, liable to drift to seaward and is frequently experienced in rivers and harbours.

(*b*) ADVECTION FOG, forming rapidly when warm moist air moves over a colder surface of land or sea.

There are also two less common types of fog:

(*c*) MIXING FOG, which forms at the boundary layer of two completely different air masses. This is also known as 'frontal fog' because it often occurs during the passage of a front.

(*d*) SEA SMOKE, which occurs when very cold air flows over relatively warm water.

Radiation Fog

The development of radiation fog depends upon the cooling of the ground at night. It is therefore a land fog, but it may drift over coastal waters with a slight wind. Very dense fogs are often formed in this manner, and those occurring in the eastern English Channel and in the Thames Estuary are mostly fogs of this type.

The cooling of the ground at night is communicated to the air in contact with it and the cooling effect is spread upwards by turbulent mixing. Since the cooling takes place at the ground, an inversion tends to develop, with the lowest temperatures on the ground; the dew-point is therefore first reached on the surface itself, and considerable moisture may be extracted from the air and deposited as dew.

On a clear night when the air is absolutely calm, the cooling is most intense close to the ground and extends upwards very slowly, and the temperature of only a shallow layer near the ground may fall below the dew-point so that fog forms within it. If there is just a little turbulence, which will be the case if there is a very light breeze, mixing occurs and the cooling spreads to higher layers, say up to 500 ft or more. Fog may then form up to that level. With stronger winds at low levels turbulence will extend through several thousand feet, and the cooling effect will be distributed through these layers so that the fall of temperature will be very small and the dew-point will not be reached at any level. Since turbulence tends to establish a dry-adiabatic lapse rate through a layer which is free of cloud initially, thereby steepening the lapse rate, the temperature will fall at the top of this layer and an inversion will frequently be found just above the layer. After turbulence has been distributing moisture upwards through the layer for some time, the condensation level will be found near the top of the layer, resulting in the formation of a layer of stratus or stratocumulus cloud just below the inversion, while no fog occurs at the surface.

From this it can be understood that the conditions favouring formation of radiation fog are:

(*a*) Large moisture content in the lowest layers.

(*b*) Little or no cloud.

(*c*) Light breeze at the surface.

(*d*) Surface of the ground initially cold and wet.

Diurnal and Seasonal Variation of Radiation Fog

The minimum night temperature occurs about dawn and the highest frequency of radiation fog occurs about an hour after sunrise. The slight increase in turbulence due to the first heat from the sun reaching the lowest layers often causes a sudden formation of fog or results in a thickening of any fog already formed in the night. Further heating by the sun is needed to clear this fog. In the British Isles, in summer, radiation fogs rarely last more than a few hours, and it is the long nights of autumn and winter which provide conditions most favourable for radiation fog. In these islands the late autumn is more subject to fogs of this type than the winter. This is because the moisture content of the air masses arriving from the Atlantic is higher in autumn than in winter, which is associated with the sea being warmer in the former season. The persistent radiation fogs, which are partly responsible for the occurrence of 'smog' in industrial areas, are formed when air stagnates over the British Isles in an anticyclone lasting for several days. In these conditions the heat loss by radiation from the lowest layers is much greater than the gain of heat from the sun through the 24 hours, and so these layers will become colder and colder. A strong inversion of temperature is soon formed at a low level, usually around 1,000 ft, which prohibits convection and traps all the industrial smoke and other forms of pollution in the air in the lowest few hundreds of feet. The only natural method by which this 'smog' is removed is by a general increase in the wind circulation. Artificial methods of reducing the nuisance are by the use of coke in place of coal whenever possible, by the abolition of the open grate and by the filtering out of solid impurities from the effluents before these are allowed to leave the factory chimney. The former involves a major change in national habits; in regard to the latter some progress has been made in reducing the quantity of solid impurities released into the atmosphere over industrial areas. In a few large cities in Great Britain and other countries smokeless zones have been instituted recently. This is a further step in the right direction although much remains to be done before a really clean atmosphere is achieved in many industrial and dockside areas.

Local Effects on Radiation Fogs

The topography and ground conditions are responsible for the local nature of radiation fog which is, however, occasionally widespread. It has a tendency to collect first in valleys, due to the greater cooling that takes place therein and to the katabatic draining of cool air into low-lying places. Although water-logged ground cools more slowly than dry ground, the effect of the increased humidity is the more important, and fogs are more likely to occur over wet ground if other conditions are suitable. Similarly, fogs are particularly likely when the sky clears at night after rain.

Under suitable conditions the fog may be just a few feet in thickness, forming a ground mist. Most radiation fogs have a depth of about 500 ft and rarely more than 1,000 ft. The upper surface is usually sharp. Owing to the fact that dust and moisture cannot be lifted through an inversion by turbulence, the clearness of the air above is usually in marked contrast to that of the air below.

Advection Fog

This type of fog occurs when warm, damp air moves over a surface cooler than its dew point. It may occur over either land or sea. Ashore, it is particularly likely when in winter, after a cold spell, a supply of milder air arrives from the sea. The air is cooled by the colder land surface over which it passes and fog is formed. If the wind is more than moderate, turbulence lifts the condensation level above the surface and low cloud forms instead of fog. Over the sea, this type of fog occurs when warm, damp air moves from the land over a colder sea, or from a region of fairly warm sea water to one of colder sea water. The frequent fogs on the Grand Banks of Newfoundland are formed when the warm damp air overlying the Gulf Stream is blown northward to the region over the cold Labrador Current. The fogs along a narrow strip of the Californian coast, which are most frequent near San Francisco, are due to the cooling of the sea breeze during its passage across the cold inshore waters of the Californian Current.

Smoke Pollution

In Britain, particularly near industrial areas, fogs are intensified by smoke pollution from industrial and domestic fires. The larger particles of soot and smoke settle easily under gravity and do not drift far, but much of the pollution is in the form of minute particles, comparable in size with the water droplets in a cloud, which may remain suspended in the air for an indefinite period.

The thickness of smoke haze depends largely on the rate at which it is dispersed through the air, and the dispersion may occur in two ways:

(*a*) Horizontally, by being transported in the wind (advection).

(*b*) Vertically, in rising currents of air (convection).

Convection currents are most effective in moving the smoke to higher levels, to be dispersed in the stronger winds aloft. Even when humidity is insufficient to produce a 'water-drop fog', smoke alone may reduce visibility below the value at which fog is defined as occurring, if the wind is light and other conditions are suitable. The foggiest regions of Great Britain are therefore in the industrial areas, where the fogs are more frequent, thicker and more persistent.

Mist, Haze, Dust* and Smoke

Mist is of similar formation to fog, but the visibility is not so seriously affected. By international agreement, visibility which is impaired, but is not less than 1,100 yards, is described as mist when the obscurity is caused by condensed water particles, and as haze when the obscurity results from smoke, dust particles or other impurities in suspension in the atmosphere.

Over the sea, a condition of haze may be caused partly by the presence of tiny salt particles in the air. Thus the difference between mist and haze is that mist is due to water particles, whereas haze is caused by solid particles.

Off large industrial centres near the shore, industrial haze from the numerous factories may be encountered some distance to seaward in certain wind conditions. The smoke from forest fires such as are experienced in the United States and Canada can similarly cause considerable haze when it drifts to seaward.

In desert or other arid regions, the visibility may be greatly reduced by dust or sand in the atmosphere. Dust or sand is raised from the ground by the wind and carried upwards to a greater or less degree according to meteorological conditions and the nature of the dust particles. This phenomenon is known as a sandstorm or a duststorm. The effects of such storms may be observed well out to sea when arid regions border the coast. Off the West African coast, dust haze

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* *See* also *Marine Observer's Handbook,* Chapter 11.

from the Sahara is experienced far to seaward during certain seasons of the year, as when the Harmattan wind is blowing.

Mixing Fog

Mixing fog is liable to occur when air streams of widely different origin meet. For example, if a cold air current meets a warm moist air current, the latter will be cooled at the boundary and fog may form there. Fog near a warm front or occlusion is quite common over the sea in temperate and high latitudes, and may be persistent in an area when a front becomes nearly stationary. Mixing fog is often known as frontal fog.

Sea Smoke

The description 'sea smoke' is given to a peculiar kind of surface mist or fog observed close to the open sea surface when the air temperature is very low. As a rule this layer of mist or fog is shallow and visibility within it is rather variable. There is some evidence that sea smoke does not form unless the air temperature is at least 16°F below the temperature of the sea surface, which explains why it is more commonly observed in the polar regions and off the east coasts of the continents than elsewhere, since cold off-shore winds are frequent there in autumn and winter. Sea smoke results from the rapidity with which cold air becomes saturated by evaporation from a relatively warm sea surface. The sea also supplies a large amount of heat, in addition to moisture, to the lowest layers of air, with the result that such a large air-sea temperature difference can only persist while strong or gale force winds continually renew the supply of cold air.

Sea smoke is most common in Arctic and Antarctic waters and in areas such as the Baltic, but it can also occur elsewhere. For example, in the Newfoundland region and the Gulf of St. Lawrence dense sea smoke (in which visibility was below 1,000 yd and which sometimes extended to a height of 5,000 ft) has been observed by ships and aircraft during the winter, with surface air temperatures below 15°F and gusty winds between W and N, often exceeding gale force. This kind of extensive deep-sea smoke is given the local name of 'ice crystal fog', and constitutes a serious hazard to navigation. As a rule sea smoke is shallow and patchy and does not impede navigation to any extent. It has been observed occasionally in this form locally in the Mediterranean, at Hong Kong and once from an ocean weather ship in the north-east Atlantic, also in the Gulf of Mexico during the occurrence of a norther (*see* page 90). Sea smoke is sometimes described as frost smoke, or Arctic sea smoke.

Visibility at Night

When making an observation of visibility at night in the open sea the distance the observer uses to describe the visibility should refer to the greatest distance which he estimates he could see and identify supposing normal daylight prevailed. For details about the determination of visibility at sea the *Marine Observer's Handbook*, Chapter 4, should be consulted.

In recent years visibility meters have been brought into use ashore for determining visibility at night. These depend for their working on using a simple device known as an 'optical wedge', which is used in conjunction with one or more lights of known brightness and at known distances from the observer. An optical wedge is a strip of glass made completely opaque at one end, the opacity decreasing uniformly to the other end where it is quite transparent. The use of the instrument depends upon the fact that the appropriate opaqueness of the position on the optical wedge which is just sufficient to obscure a light of known intensity at a known distance, when it is viewed through this part of the wedge, can be readily calculated, assuming that the atmosphere between is perfectly transparent. When the observer makes an observation in practice, he finds that a position on the wedge, where the absorption due to it is rather less than at the calculated position, is sufficient to render the distant light invisible to him. From the difference between these two positions on the wedge he knows that the additional opacity which is not supplied by the wedge is provided by the length of air between the light and the observer; and in this way, provided certain precautions are observed such as keeping the wedge and other glass surfaces absolutely clean and the voltage of the lamps constant, an accurate determination of visibility can be made.

The night-time visibility as defined for meteorological purposes does not give a reliable estimate of the distance at which a lighthouse will be visible to a ship at sea, for the reason that this distance will depend upon the brightness (candle power) of the lantern. The amount of obscurity in the atmosphere is also liable to vary as between different levels, so that the apparent brightness of the lantern on a tall lighthouse to an observer on a ship is by no means a reliable guide to the true visibility close to sea level. The possibility of a light being obscured by low cloud, which may be formed from a fog bank recently lifted from the sea surface by a change of wind, should always be borne in mind by the navigator when making a landfall.

Exceptionally Good Visibility

Occasionally visibility is exceptionally good and in such cases mountain peaks over 100 miles distant may be seen with clarity. In these cases the air is very pure. These conditions, when they occur, are most common in air of polar origin or in air which has spent a long time over an ocean in tropical regions.

Audibility of Sound Signals in Fog

It is rightly said in the books of seamanship that 'sound is conveyed in a capricious manner through the atmosphere'. It is important to realize this when considering the distance at which one may expect to hear a sound signal in fog. In calm, or relatively calm, weather the strength at which sound signals are heard from a stated distance in fog may vary from day to day, or on a given day the signal audibility may vary on different azimuths from the observer. Sound waves may at times be reflected or refracted, and hence attenuated, due to the presence of the excessive number of water particles present in a foggy atmosphere. The echo which is often heard from a ship's whistle in fog is evidence of this. In a breeze it is fairly obvious that sound signals originating from a source situated to leeward of the observer, will not normally be heard at as great a distance as from a source to windward. When the wind is blowing at all hard it is often difficult to hear fog signals owing to the noise of wind and sea. The mariner needs to be constantly on his guard when estimating distance from an object in foggy weather by the strength of its fog signal or its apparent azimuth.

Use of Radar in Fog

When using radar for navigation in conditions of low visibility, the navigator should bear in mind the meteorological factors which affect the performance of the instrument. A reduction in the expected range may be because of conditions favourable to sub-refraction or because the radar impulses are attenuated due to absorption by water particles in the atmosphere. On the other hand, ranges may be considerably greater than expected if conditions are favourable for super-refraction. Standard atmosphere conditions may be defined broadly as follows:

(*a*) Barometric pressure at sea level 1013 mb, decreasing with height at a rate of 36 mb per 1,000 ft.

(*b*) Temperature at sea level 59°F, decreasing with height at a rate of 3·6°F per 1,000 ft.

(*c*) Relative humidity 60% (remaining constant with height).

Under such conditions the radar range should be normal, i.e. it should somewhat exceed that of the optical horizon. If the temperature decrease with height is more than standard, or if humidity increases with height, then sub-refraction may be expected and the radar range may be considerably reduced. If the temperature decrease with height is less than the standard, or if humidity decreases in height, then super-refraction may be expected and the radar range may be considerably increased. Another form of super-refraction known as 'ducting' is also experienced due to irregular change in the temperature–humidity gradient in height between the observer and the target, and this may give excessive radar ranges. It is perhaps unfortunate that super-refraction is most frequent when light winds carry warm dry air across a relatively cool sea, that is when the weather is clear. Although the observer aboard a ship cannot be sure when either condition is present without an accurate knowledge of the vertical gradients of temperature and humidity, he can at least obtain some indication as to when conditions are very favourable for sub-refraction or super-refraction by taking readings of dry-bulb and sea temperature. When the air temperature is considerably below the sea temperature (by 10°F or more) it is likely that sub-refraction is present. Alternatively, when the air is 10°F or more warmer than the sea it is almost certain that super-refraction conditions exist. In both cases this is only true with light and moderate winds when turbulence is not marked. It has also to be remembered that the radar range regarded as normal for one region would be abnormal for another region, and that the range in many localities varies between the seasons.

Reduction of radar range by attenuation may be caused by the presence of fog, rain, ice particles, hail or snow due to the concentration of water particles therein. Generally the larger the water particles and the closer they are concentrated the greater the attenuation, so that the effect in heavy rainfall is probably greatest and that due to fog is least. The effect of snow and hail can normally be expected to be less than that of rain at an equivalent rate of precipitation. Thus, in the conditions which are the greatest menace to navigation (fog and snow) the probability is that the attenuation will not be as great as in heavy rain. Some attenuation may be experienced due to the dust, sand or smoke particles present in haze. The meteorological aspects of the use of radar at sea are more fully dealt with in Chapters 5, 6 and 7 of the book *The use of Radar at Sea*, published by the Institute of Navigation.

PART II. CLIMATOLOGY

CHAPTER 6

METEOROLOGICAL CHARTS OF THE OCEANS

Use made of Ships' Meteorological Observations

Meteorological observations, whether taken on land or at sea, serve several purposes. When transmitted by radio to a central meteorological office and there plotted on a synoptic chart, they provide an up-to-date picture of the weather over a large area. From a study of such a chart, together with other synoptic charts plotted several hours earlier, forecasts can be made of the weather for the following 12 to 24 hours. As much of western Europe's weather comes from across the Atlantic, the importance of ships' observations for synoptic purposes in that area is obvious. The same is true for other parts of the world. As about three-fourths of the earth's surface is water, synoptic charts would be very incomplete without the help afforded by those ships' officers who voluntarily provide information about the weather at sea.

These observations are also used statistically. If enough observations can be collected from one place, or from a limited area, they will show, among other things, the most frequent type of weather, or within what limits any element such as temperature, pressure or wind may be expected to vary, at any time of the year. Such information is of value in cases where daily forecasts are unobtainable or inadequate; as, for example, when it is required to know the most favourable time of year for carrying out a difficult operation such as towing a floating dock or a disabled ship to a distant port. Statistics of sea and air temperature and humidity can be valuable in dealing with practical problems such as refrigeration of perishable cargoes at sea, and prevention of damage to cargoes in transit due to rust, condensation and similar agencies.

Meteorological observations, both in the form of statistical summaries and when plotted on synoptic charts, are also used for research into the causes of changes in weather. Until these causes are more fully understood, no great advance in forecasting technique is probable.

Some observations, such as barometric tendency and characteristic and cloud types, are chiefly used for forecasting; others, such as wind, temperature and humidity observations, are very necessary for forecasting and are used also for statistical and research purposes.

There is one essential difference between land and marine climatology. On land there are regular series of observations, made several times a day over a long period of years at fixed locations, from which a complete picture of the climatology can be built up. At sea the situation is different; observations are only obtainable from ships, and even on the main shipping routes the chance of getting many observations from one spot, or even from a comparatively small area, say one square mile, is only slight. The exceptions are in the North Atlantic and North Pacific oceans, where the existence of ocean weather stations since 1947 has provided a regular series of observations at fixed points in the

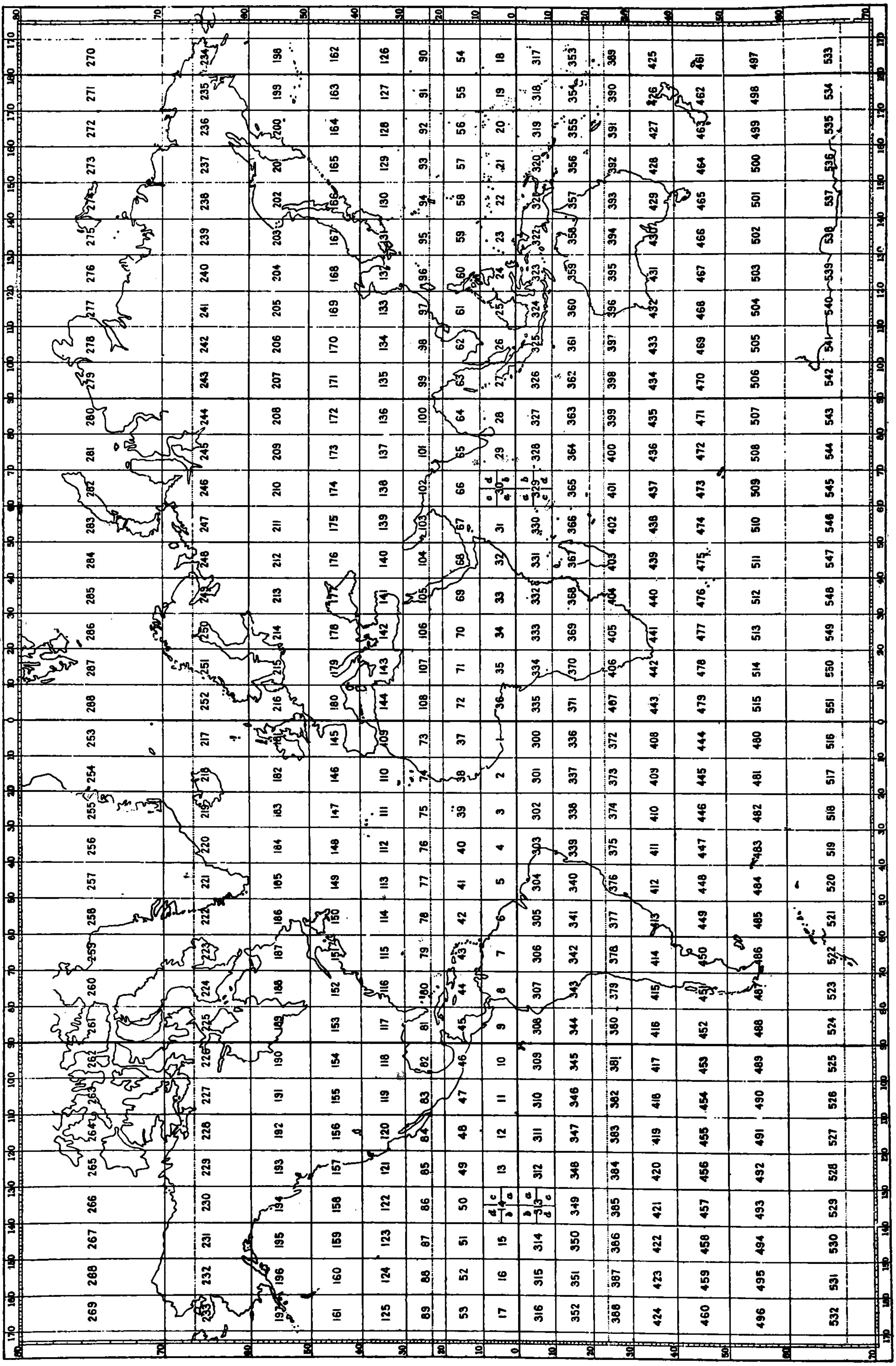

Fig. 6.1. Marsden chart of the world

ocean. It is necessary, therefore, to divide up the ocean into relatively large areas, at the very least into 1° 'squares', that is, areas bounded by two meridians and by two parallels 1° apart, while 2°, 5° and even 10° squares are also used. All observations taken in such a square are grouped together as if they had been taken in the same spot.

Marsden System of defining Ocean Areas

In the Marsden system, used by the Meteorological Office for marine climatology, the surface of the earth is divided up into 10° 'squares' bounded by meridians and parallels at intervals of 10° (*see* Fig. 6.1). These areas are known as Marsden squares, named after a former Secretary to the Admiralty, who introduced this system in 1831. As can be seen from the diagram, in both the northern and southern hemispheres the numbers start at the intersection of the Greenwich meridian and the Equator. In the northern hemisphere the first number (to the west of the Greenwich meridian) is square 1, in the southern hemisphere square 300.

Each Marsden square is further sub-divided into four 5° squares and 100 1° squares. The scheme of labels (number or letter) for these smaller squares depends on whether the longitude is west or east of the Greenwich meridian, the latitude north or south. Fig. 6.2 shows the scheme for each of the four quadrants, NE, NW, SE, SW.

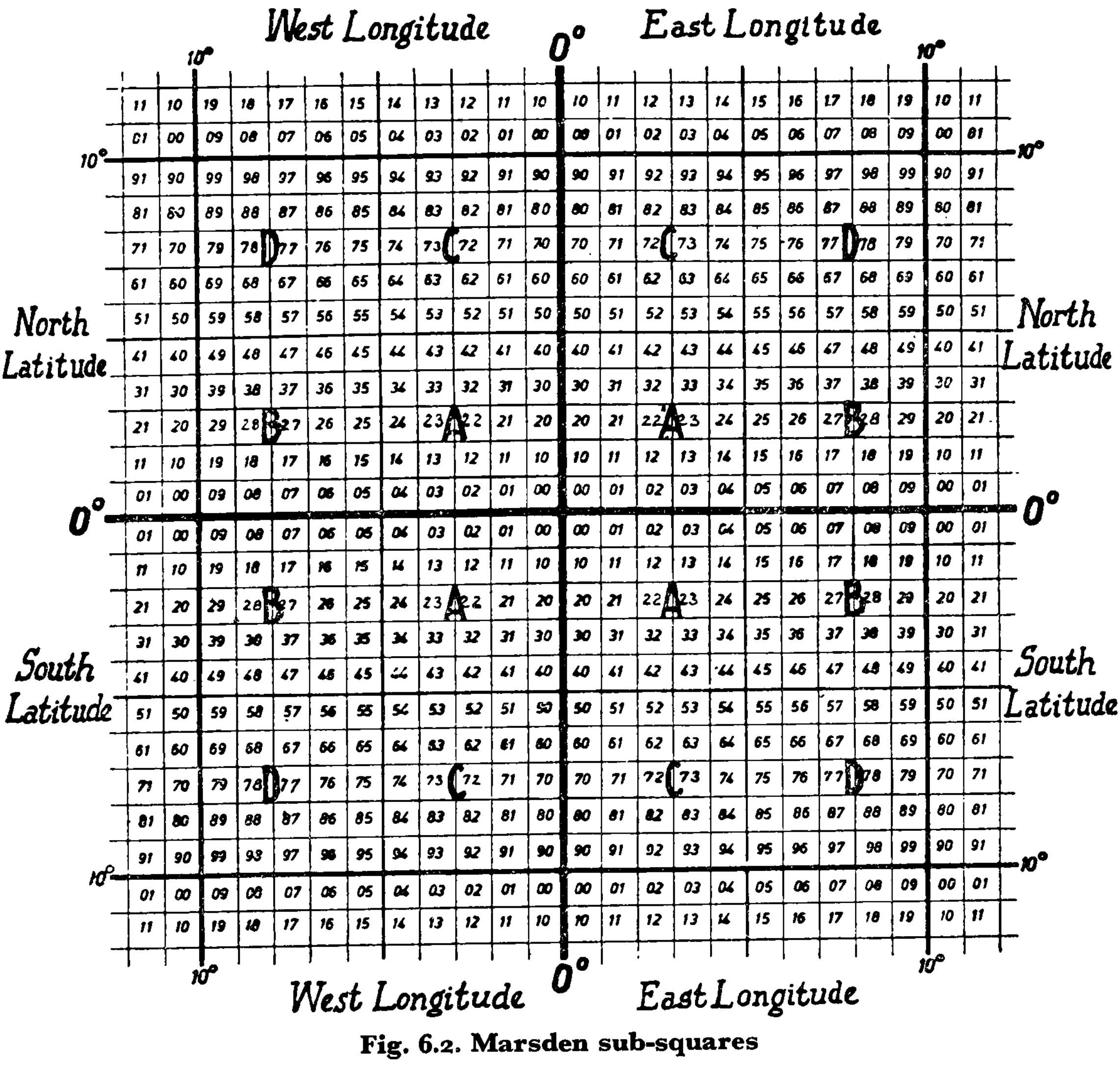

Fig. 6.2. Marsden sub-squares

The 1° square in which a ship happens to be can be quickly deduced from its position as given in the log. The unit figures in the number of degrees of latitude and longitude, respectively, give the number of the 1° square. For example, if a ship's position is given as 47° 42'N, 33° 23'W, from the Marsden chart (Fig. 6.1) it is seen to be in square 148, and from the unit figures of the degrees of latitude and longitude that it is in the 1° square 73 (Fig. 6.2). If required, the 5° square can then be deduced from the 1° square. The advantage of the Marsden square system is that it provides a shorthand method whereby a ship's position can be rapidly described to the nearest 1° square, which is sufficiently close to its true position for all normal maritime meteorological purposes.

The fact that observations from a fairly large area are considered representative of one spot is not such a disadvantage in the open sea as it would be on land, for climate does not generally change as rapidly with position over the sea as it does over land. Near the coast, however, and at the boundaries separating ocean currents of different temperatures, the climate may change rapidly over short distances, and observations in these regions may require special treatment.

There is no exact regularity, such as there is on land, about observations at sea, except in such special cases as those from ocean weather ships when on station and from lightvessels. As a ship passes through a particular square, a series of observations is obtained at successive intervals of 6 or perhaps 12 hours, but that square may contain no further observations during the month in question for days, weeks or even years. In effect, sample observations of the weather are made. These samples are not even independent ones for consecutive observations, at intervals of a few hours, are liable to give much the same type of weather. However, when enough of these samples are obtained, we should have a reasonable approximation to the statistical results which would be shown by regular series of observations.

Another difficulty is that a great proportion of ships' observations are made along the principal shipping routes of the world, so that the number of observations in neighbouring Marsden squares may vary considerably. There are still large areas of the ocean where there are squares containing very few observations. Even where there are a few observations, these may all have been taken in one, two or three years only; in such a case they can only give a very approximate representation of the various climatic elements in that square.

Ocean Weather Stations

The existence of ocean weather stations in the North Atlantic is important for marine climatology. These ocean weather stations are comparable with first-class land stations in the regularity and extent of the observations, both surface and upper air, taken at a 'fixed' location. After a few years there should be available from these ships much data, which will supplement those received from other ships, and also provide urgently needed information both for the surface and the upper air, in a form in which it can be used to investigate problems of short- and long-range (seasonal) forecasting in Great Britain and northern Europe.

The presence of these ocean weather stations by no means lessens the value of regular weather observations by other ships. The expense of providing and maintaining ocean weather ships is considerable and they are necessarily few in number. From a meteorological standpoint it is as though some new islands had been located in the North Atlantic, but information from the large areas between them is just as necessary as before and must be obtained from merchant ships in order to obtain a comprehensive synoptic picture.

Manipulation of Data

When observations are received at the Marine Division of the Meteorological Office, they are scrutinized and classified and then punched on Hollerith cards, so that as much as possible of the work can be done by machine methods. The observations are then grouped by months and in areas of appropriate size. Temperature observations are usually grouped in 2° squares and pressure, wind, cloud and weather in 5° squares.

The cards are then sorted by mechanical means and the results are arranged either in the form of monthly means, i.e. mean temperatures, mean maximum and minimum temperatures, mean pressures, etc., or in monthly percentage frequencies. The percentage frequency of fog, for example, in any square is expressed in the form:

$$\frac{\text{Number of occasions of fog}}{\text{Total number of observations}} \times 100.$$

When these results have been calculated, they are portrayed by means of isopleths on charts of the region concerned. An isopleth is a line drawn on a chart so as to pass through all the points at which any element has the same value. Examples of isopleths are isobars, joining points at which the pressure is the same; isotherms, joining points having the same temperature; isonephs, for cloudiness; isohyets, for rainfall; and isallobars, which join points having the same barometric tendency.

Portrayal of Distribution of Vector Quantities

Wind and current possess two properties, those of magnitude and direction, which are known as vector quantities. The isopleth method is not suitable for portraying them, except when only one constituent is to be shown and the other is ignored: for instance, to show the frequency of gales, irrespective of direction.

The most important vector quantity in marine climatology is the wind. Wind observations are usually summarised in the form of wind roses. These are diagrams showing, for each area, the percentage frequency of winds from each direction by means of arrows, the length of each arrow being proportional to the percentage frequency. Every arrow is also divided into segments, each segment representing winds between certain limits of strength.

Fig. 6.3 shows the wind rose used in the Marine Division of the Meteorological

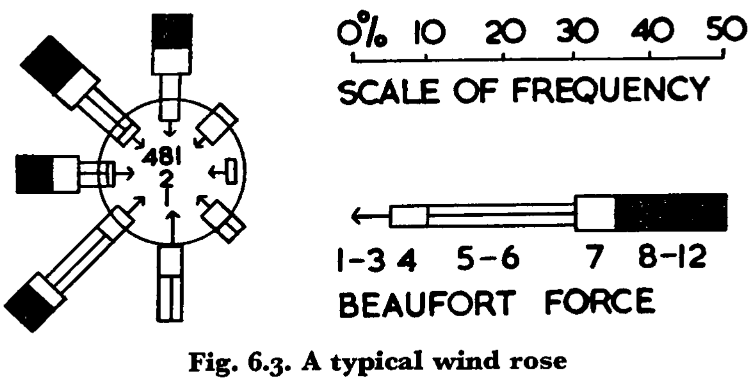

Fig. 6.3. A typical wind rose

Office. The directions shown are the eight compass points, N, NE, E, etc., all directions between NNE and ENE being classed as NE and so on. Within 30° of the equator, however, the 16 points N, NNE, NE, ENE, etc., are used, because the trade winds and monsoons tend to be constant in direction and the use of eight points only might obscure this constancy. For each direction the percentage frequencies of Beaufort force groups 1–3, 4, 5–6, 7, 8–12 are shown by the lengths of segments of the arrow. The thickness and shading of the segments indicate which group of wind forces they represent.

From the wind rose a rough estimate can be made of the direction from which the wind most frequently blows. It is, however, better to show this explicitly on a separate chart, in the form of arrows, thus indicating the direction of the prevailing or predominant winds.

Another method of summarising wind data is to show the vector mean of all the winds observed in a given area. This is of value in showing the general flow lines of the main air currents, and for investigations into the relation between these and ocean currents. Vector mean winds, however, give no idea of the mean wind strength. For example, if during half the month there are easterly winds of force 5, and for the other half westerly winds of the same force, these two will cancel out in the vector mean wind, which will therefore be force 0, yet the average wind during the month, irrespective of direction, is force 5. This is an extreme case, although it is usually found that the strength of the vector mean wind is appreciably below the wind strength averaged irrespective of direction.

Marine Meteorological Atlases

In the Marine Division atlases have been prepared covering the Atlantic, the Indian Ocean, the western and the eastern Pacific. These contain monthly charts showing isotherms of mean maximum and minimum temperatures for both the air and the sea surface; isotherms of mean air-sea temperature differences; isopleths showing the frequency of precipitation, snow, lightning, fog, poor visibility and light, moderate and heavy cloud. Isobars and isonephs of mean cloud are also shown. The wind data include not only wind roses, but predominant and mean vector wind arrows and isopleths of the frequencies of gales and of light or moderate winds. Swell roses are also included. These atlases are supplied free to voluntary observing ships on request.

Some of the earliest examples of climatic charts of the oceans were those prepared by Commander Maury of the United States Navy in about 1858, showing winds and currents. With the aid of these charts, the sailing-ship passage to Australia, which had previously taken on an average 124 days, was reduced to an average of 97 days.

CHAPTER 7

THE WIND AND PRESSURE SYSTEMS OF THE OCEANS

Main Features

In Chapter 2 the main features of the pressure systems shown on synoptic charts were described. A synoptic chart portrays in international symbols the meteorological observations in the area which it represents, including the distribution of pressure, for one particular occasion. Any completed working chart, such as one of those used for plotting surface observations at, say, any important forecasting office in Great Britain (and which is on a small enough scale to show most of the North Atlantic, north-east Canada and U.S.A., besides Europe, North Africa and western Asia) makes the number and complexity of the pressure systems in existence at any one time obvious almost at a glance. The study of any series of synoptic charts suggests that an attempt to prepare a chart showing the average conditions in the same area through the year or through a particular month would not yield results of any value. However, this is fortunately not the case. In the Marine Division of the Meteorological Office charts have been prepared which show monthly mean values of pressure mainly for 2° sub-squares and of wind for 5° Marsden sub-squares covering a large proportion of the surface of the oceans.* The observations upon which these charts were based have been made by selected ships since 1855, and were later punched onto Hollerith cards. These charts show that there are large and well-defined regions of high and low pressure and associated flow of air which persist from month to month throughout the year, or during a considerable part of it. In general the geographical distribution of these areas of high and low pressure is fairly systematic and does not vary much between consecutive months. However, there are some large differences between the pressure patterns and prevailing winds shown on these charts for the summer and winter months, notably over the continents in the northern hemisphere. These prevailing winds, and the vertical air movements which accompany them, comprise the general circulation of the atmosphere. Figs. 7.1 and 7.2 show the average pressure at mean sea level, over the oceans, in January and July. When these charts are compared with similar charts showing the vector mean† surface winds (*see* Figs. 7.3 and 7.4), it can be seen that the average distribution of pressure is related to these winds in accordance with Buys Ballot's law. However, care is needed in the interpretation of vector mean winds. Some extremely windy regions of the world, such as the neighbourhood of Cape Horn, show rather low values of vector mean wind in comparison with the mean wind speed of the area. This is because in such regions the direction of the wind is variable, and strong winds occur with nearly equal frequency from several directions. In the doldrums vector mean wind values are very small, since light winds are common there and can blow from several directions. If the wind in a locality always blew from the same direction, the vector mean

* Similar climatological maps have been prepared by other authorities for land and sea areas.
† *See* footnote on page 84 for explanation of 'vector mean wind'.

wind value would be equal to the mean wind speed in that locality. The steadiness of direction of the trade winds is well shown by the fact that vector mean wind values computed in the trade wind regions approach the local mean wind speeds more closely than in any other areas.

General Circulation

Before we look at the average monthly pressure distribution it will be a help to remind ourselves of some factors already described which have a great influence in shaping the general circulation as we know it.

In the first place there is the great difference between the amount of the sun's heat which is received on the tropical and on the polar regions. This difference is largely responsible for the highest surface temperatures being found mainly in tropical regions and for the lowest surface temperatures occurring in or adjacent to the polar regions. This excess of solar heat reaching the earth's surface in the tropics leads to convection in the equatorial regions on a greater scale than anywhere else on the earth. The widespread raising of portions of air from the surface to high levels in the troposphere, brought about by this convection, is compensated by a general flow of air at low levels towards the equator and a corresponding poleward flow away from the equator at high levels. From these simple considerations it might be expected that a belt of low pressure would extend around the equatorial regions of the earth, as it does in reality.

However it is now realised that many other factors, such as convergence and divergence of air at different levels, will have to be taken into account before we can fully account for the existence of the equatorial low pressures and many other features of the general circulation.

So far we have neglected the effect of the earth's rotation (geostrophic force). When this is taken into account we can see why the low-level winds in the tropics are deflected from northerly to become north-easterly in the northern hemisphere and from southerly to become south-easterly in the southern hemisphere. In a similar manner the flow of air at higher levels in both hemispheres, which is directed away from the equator initially, is subject to the geostrophic force, the magnitude of which is dependent upon the sine of the latitude as described in Chapter 3. As this air moves to higher latitudes the geostrophic force is increasing and deflects the air so that it blows from a direction which becomes increasingly westerly, i.e. along the parallels of latitude. A little consideration shows that there is likely to be an accumulation of air in these latitudes from this cause on both sides of the equator. At the same time a part of this air-flow away from the equator at higher levels descends slowly in latitudes of about 30° into the areas occupied by the semi-permanent subtropical anticyclones. This air is steadily losing heat by radiation as it moves to higher latitudes, and this factor also helps to bring about gradual descent (subsidence) of the air on a large scale. As we have seen earlier, large-scale descent of air is usually associated with anticyclones, which in the general circulation are frequently found over the oceans in subtropical latitudes. On the poleward flanks of these anticyclones currents of air having a general westerly direction are found at all levels in the troposphere.

In the polar regions for most of the year a continual loss of heat by radiation takes place from the atmosphere at all levels. This results in subsidence of air at low levels, leading to an average outflow of air from the polar regions near the surface; hence north-easterly or easterly winds are common over the Arctic regions as a whole, and similarly south-easterly or easterly winds are frequent over the oceans around Antarctica.

Broadly speaking, therefore, in temperate latitudes cold and relatively dry winds from an easterly direction lie to polewards of mainly moist and warm westerly winds. As we shall find in a later chapter, these circumstances are most favourable for the development of the depressions of temperate latitudes. The warmer westerlies meet and ascend over the colder easterlies at a surface of separation, the line where this surface of separation makes contact with the land or ocean nowadays being known as the POLAR FRONT. Depressions are initiated on bends or 'waves' in this front. These ideas originated about the time of the First World War by the Norwegian school of meteorologists. By working them out in greater detail meteorologists were enabled to gain a much better insight into the nature and behaviour of temperate latitude depressions than had been possible up to that time.

Surface Circulation

Before describing the main features of the surface mean circulation there are two other factors to be taken into account. Firstly, the seasonal variation in the sun's declination leads us to expect that the belt of highest temperature would 'follow the overhead sun' and thus move into the summer hemisphere, and implies that the equatorial low-pressure belt would undergo a similar seasonal migration. This occurs in reality, and there is also a less obvious seasonal movement over a few degrees of latitude of most of the main pressure belts and centres of the surface mean circulation, which can be identified on most of the charts of monthly mean pressures over the world. Secondly, the uneven distribution of land and sea over the earth's surface makes the pattern of the surface mean circulation less simple than the pattern we might expect to find on an earth whose surface consisted entirely of sea or of land. Broadly speaking, in the lower (subtropical) latitudes, areas of high pressure are more pronounced over the sea than over the land, particularly in the summer hemisphere; in middle and high latitudes over the continents there is high pressure in winter but relatively low pressure ('monsoon' lows) in summer. In temperate latitudes a low-pressure area is found over each ocean during the whole year, though the systems in the winter hemisphere are more intense as a rule. In brief, areas of high pressure in the surface mean circulation are more readily developed over surfaces which are cool or cold for their latitude, while low-pressure areas in temperate latitudes are similarly found over sea surfaces which, besides providing the moisture needed for energising depressions at all seasons, are also warm for their latitude in the winter months.

We can now examine the surface mean circulation of the atmosphere over the earth's surface as a whole (Figs. 7.1 and 7.2). Among the systems which persist as more or less prominent features throughout the year there is first the equatorial trough or belt of low pressure, in which pressure gradients are mostly weak and in which no well-marked centres of low pressure show up over any ocean. The subtropical belts of high pressure, particularly the anticyclonic centres over the oceans, and the low-pressure areas in temperate latitudes of the North Atlantic and North Pacific, besides the Southern Ocean, also show up well at all seasons.

A few major parts of the surface mean circulation are developed only during

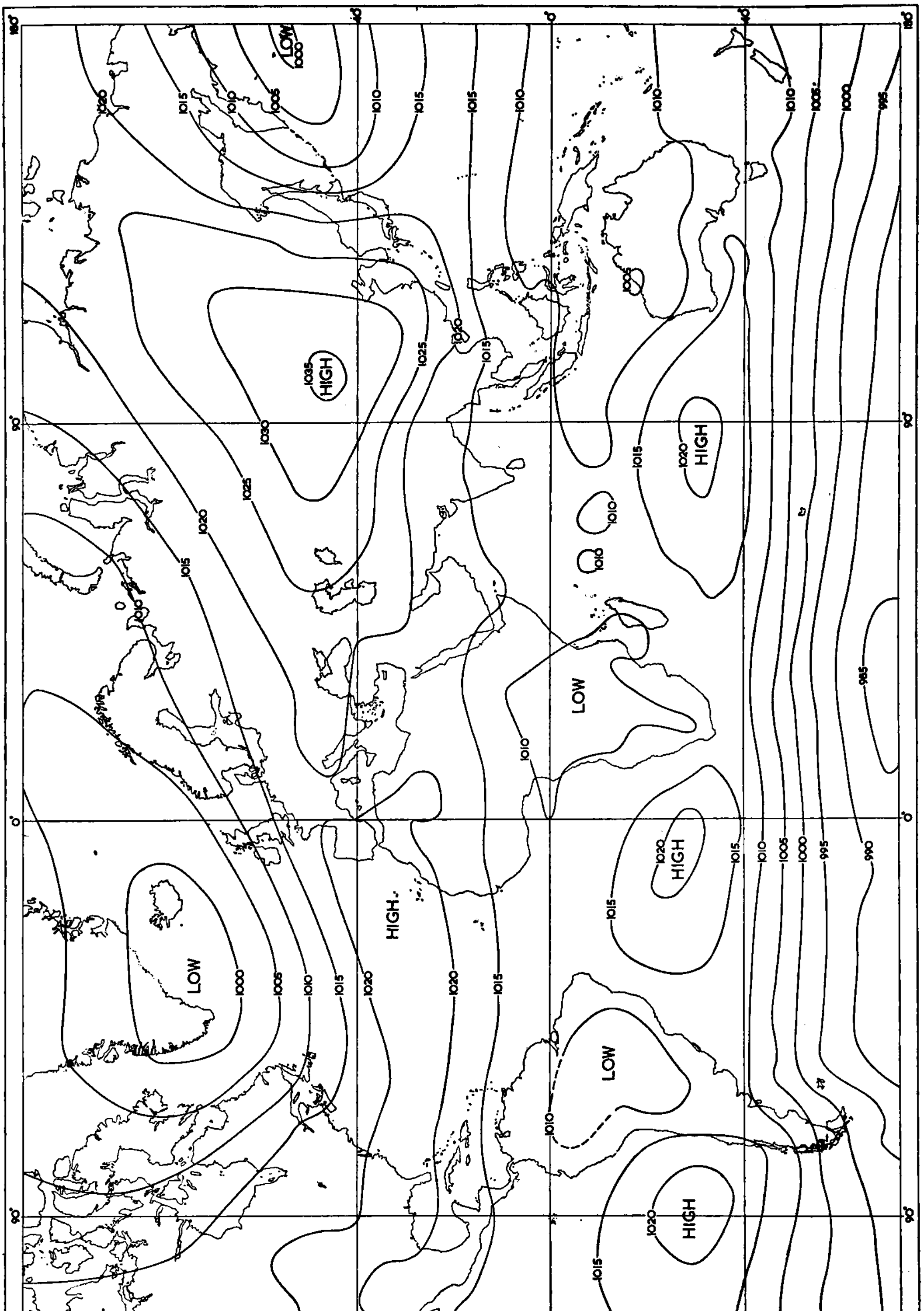

Fig. 7.1. Mean pressure, January

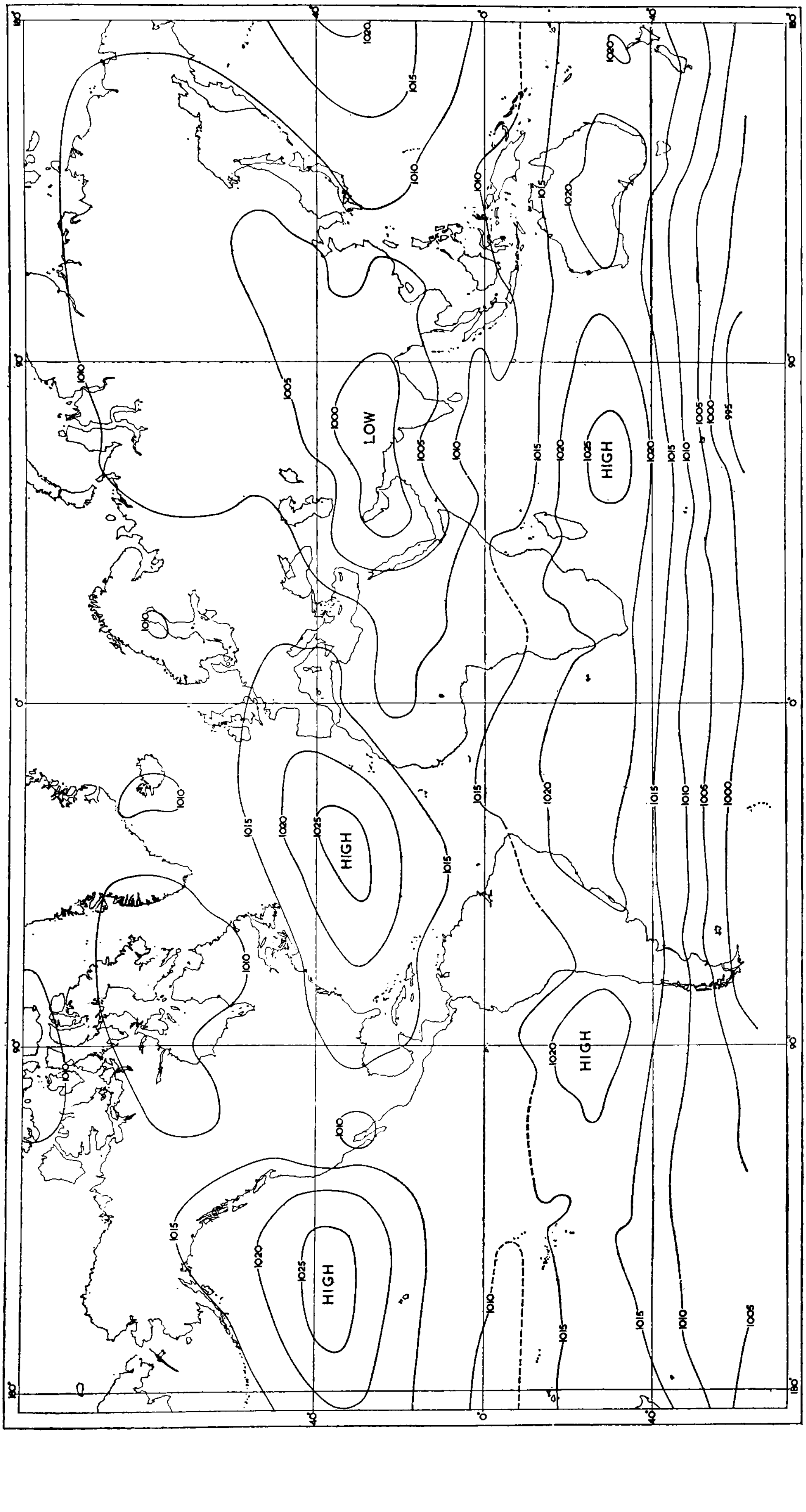

Fig. 7.2. Mean pressure, July

the summer or winter months in particular regions. Chief of these in winter are the large and intense high-pressure area found over Eurasia and the rather smaller and less intense high-pressure area found over western North America. The high pressures over Asia and North America disappear in the summer. Other features peculiar to the summer hemisphere are the monsoonal low pressures found over the continents in subtropical and lower latitudes. Of these, the large area with its lowest pressure over south-west Asia is most notable both as regards intensity and extent, and for the modification it makes to the wind circulation and climatic regimes of a large area of Asia. The corresponding low pressures formed over the south-western United States, in southern Central Africa and in South America in their respective summer months, are important features of the surface mean circulation of those regions.

The reader who wishes to study the seasonal changes in the surface mean circulation in greater detail, is recommended to examine the series of charts for each month given in the Marine Meteorological Atlases (*see* page 76). He will learn from these that the changes in the distribution of pressure between, for example, the northern hemisphere winter and summer, mainly result from two factors. One is a migration in latitude of the belts of equatorial low pressure and subtropical high pressure (following the sun's change of declination but through a much smaller range of latitude); the other is the disappearance of the winter high pressures over the continents and their replacement in summer by areas of monsoonal low pressure at a rather lower latitude than that of the winter continental high pressures. A study of these atlases will also assist in understanding the seasonal changes of wind direction, and the incidence of rainy and dry seasons in tropical and subtropical regions.

At the same time too much stress should not be laid upon the physical significance and reality of the surface mean circulation. The pressure distribution given on the charts described above is essentially the same as would be found by averaging the pressures shown on a very large number of synoptic charts, and shows nearly continuous belts of high and low pressure around the circumference of the world. But all synoptic charts referring to a real weather situation show closed isobars surrounding anticyclones and depressions of finite size, usually less than 2,000 miles in diameter. In contrast with the charts of mean surface pressures (Figs. 7.1 and 7.2), synoptic charts also show breaks in the subtropical high-pressure belts, through which outbreaks of cold air are able to reach the trade wind belt, and thereby supply the 'raw material' for the equatorial convection processes.

Pressure Distribution

Southern Hemisphere

The southern hemisphere will be considered first, since so large an area of it is covered with water that the general circulation there is simpler and more closely in accord with the idealized general circulation on a uniform earth, which has been described above.

At about 30°s a belt of high pressure goes right round the earth but changes its latitude with the sun, being some 5° further south in January than in July. Within this belt are semi-permanent areas of even higher pressure, the most prominent of which are in the South Atlantic, the South Indian Ocean and on the east side of the South Pacific, while during the southern winter there is also a well-developed seasonal high over Australia, which extends into the western South Pacific.

South of this belt the pressure falls rapidly. Observations in high latitudes are scanty, but suggest that a belt of low pressure, situated between 60° and 70°S, encircles the earth and that further south pressure rises again.

North of the southern subtropical high-pressure belt the pressure decreases, till the equatorial low-pressure belt is reached. In the Atlantic and Pacific this persists throughout the year with only minor changes; but in the Indian Ocean, during the SW monsoon (June–September), it is very much modified. At this season, owing to the effect of the sun's heat on the extensive land masses of Asia, the pressure falls fairly steadily from the subtropical high-pressure belt at about 30°S till it reaches its lowest values over northern India, Arabia and the adjoining countries. A similar effect, though on a smaller scale, occurs in the western hemisphere, where the equatorial low-pressure belt in summer protrudes northward across the equator into the western part of the American continent.

In the Atlantic and Pacific the equatorial low-pressure belt moves north and south with the sun, but only slightly, the movement in latitude being about 5° to 10°.

Northern Hemisphere

In the northern hemisphere, north of this equatorial low-pressure belt, there is another subtropical belt of high pressure. Because of the larger proportion of land in the northern hemisphere, this belt is more irregular and less continuous and its seasonal changes are much more pronounced than is the case with the corresponding belt in the south. In the Atlantic and Pacific this belt lies mainly in the subtropical regions all the year, with a mean latitude of about 30°N, and minor movements north or south as it follows the sun. Over the continents, however, seasonal changes are much more prominent. During the northern winter this high-pressure belt expands into two huge areas, stretching to the northward over the continents of Asia and North America, thus giving rise to the NE monsoon. Note how, in Fig. 7.1, the isobars in the North Atlantic and North Pacific bend sharply to the north as they approach both the east and west coasts. In the northern summer the subtropical belt of high pressure breaks up into two high-pressure areas in the North Atlantic and North Pacific, separated by the deep monsoon low which forms over Asia and by the less marked seasonal low which forms over the west of North America (*see* Fig. 7.2).

North of the main portion of this high-pressure belt there are, in winter, two well-defined semi-permanent low-pressure areas, known as the Icelandic low and the Aleutian low. These are separated from one another by the regions of high pressure over Asia and North America. The Icelandic low, though centred near Iceland, extends to the NE into the Barents Sea, while the Aleutian low lies near the Aleutian Islands. In summer the Aleutian low almost disappears; in its place there is merely a trough of low pressure over north-east Siberia which is an extension of the Asiatic monsoon low, while the Icelandic low is much less marked.

Observations in north polar regions are few. The data available suggest that in winter an area of high pressure lies north of Arctic Canada, which connects the high-pressure areas over north-west America and central Asia. This forms a col, with the trough extending NE across the Barents Sea from the Icelandic low, together with the Aleutian low lying on either side of it. In summer the pressure in polar regions is probably a little lower than in the neighbouring regions of the north temperate zones, particularly in the regions between the Norwegian Sea, Spitzbergen and Bering Strait.

General Remarks on Pressure Distributions

The features of the pressure distribution described above appear only as the result of averaging large numbers of observations. In tropical and subtropical regions, variations from the average are slight, except on the comparatively infrequent occasions when tropical cyclones develop, so that the pressure distribution at any time does not usually vary much from the average. Elsewhere this is not true, particularly in temperate latitudes where depressions and anticyclones develop and travel along a variety of tracks (although these are mainly from a westerly point to an easterly point) with great frequency. In these regions barometric pressure is frequently varying and the pressure distribution at any particular time is often completely different from that shown on the chart of mean pressure for that time of year; this is particularly true of the regions occupied by the Icelandic and Aleutian lows.

Wind Systems

Charts giving the prevailing or vector mean winds* over the earth's surface during any month generally show a close relation to the pressure distribution, particularly when the predominant direction of the wind is well marked, or its vector mean is relatively high. Figs. 7.3 and 7.4 show the vector means of the ocean winds for January and July. The length of the arrow is proportional to the vector mean, and hence to the mean rate of travel of the air mass near the sea surface.

Intertropical Convergence Zone or Doldrums

The belt of separation in the region of the doldrums between the various air currents whose origins are respectively in the northern and southern hemispheres was formerly known to meteorologists as the intertropical front, but this term has fallen into disuse, largely because it has been found that there is little resemblance between the so-called fronts of the tropics and those of temperate regions. The term INTERTROPICAL CONVERGENCE ZONE is now used, since it is generally agreed that the belt resembles a line of convergence, and this is true both on individual synoptic charts of equatorial and tropical regions and of its mean position for a month or season, which can readily be traced across the greater part of the oceans on charts of mean winds. It is shown as a pecked line in Figs. 7.3 and 7.4. It passes through the mean position of the doldrums and follows the sun, being generally 10° to 20° further north in July than in January. In January it can be traced across each ocean. In July, however, the Pacific section of this convergence zone moves north as it approaches the eastern shores of Asia, and it is not easily identified in the China Seas or Indian Ocean, where the only winds to be found at this season originate in the southern hemisphere (*see* following paragraph). Those parts of the region immediately adjacent to the convergence zone, and which are also occupied by the trade winds of both hemispheres, are naturally regions of frequent rainfall.

* The vector mean wind of a number of separate wind observations is obtained by first resolving each observed wind into its components (northern and eastern positive, southern and western negative), finding the algebraic sum of these and hence the mean value of each component. From these mean components the vector mean wind is found by simple trigonometry or the use of tables. The vector mean wind thus takes account of the directions, as well as the velocities, of all the observed winds.

On the eastern sides of the Atlantic and Pacific Oceans the NE and SE trades die away as they approach one another, and the two air streams are often separated by a region of calms and light winds which is known as the doldrums. Further west the two trade winds blow more nearly parallel and may finally blow side by side from a near easterly direction. The doldrum areas also follow the north and south movement of the sun in declination but with a much smaller range in latitude, and they are not always coincident with the position of the intertropical convergence zone. In the Atlantic the doldrums remain north of the equator throughout the year.

Trade Winds

The trade winds of the Atlantic, eastern Pacific and southern Indian Ocean show up clearly on both charts as airstreams with a high mean rate of travel. In each of these oceans they blow more or less constantly throughout the year from about lat. 30° towards the equator. In each case they extend furthest from the equator on the east side of the relevant high pressure area and they blow round that area towards the doldrums or equator changing direction from north to north-east (S to SE) to ENE (ESE) or even to east except in the South Indian Ocean in the northern summer when the south-east trade-winds cross the equator to become the SW monsoon. The trade wind areas tend to follow the north and south movement of the sun in declination; in the southern hemisphere the movement is small, but in the northern hemisphere in the Atlantic and Pacific Oceans the zone of highest trade wind speeds moves through 8° to 10°. In the Atlantic the mean wind speed of the trades is about 13 to 15 knots, the higher value occurring in the NE trade. The highest mean value anywhere (18 knots) is found in the SE trades of the Indian Ocean. In general the trades in each hemisphere blow most strongly at the end of the winter.

Monsoons

In the northern parts of the Indian Ocean and western Pacific the trade wind system disappears, and we find instead a system of seasonal winds known as the monsoons. In winter (northern hemisphere) these monsoon winds blow from a north-easterly direction under the influence of the Asiatic high centred over Siberia. As they cross the equator they back, and become much less constant NW to W winds, and meet the SE trades at about 10°S over the Indian Ocean in January. In summer (northern hemisphere) the Asiatic high disappears, and is replaced by a deep and extensive low, in which the lowest pressure is found over south-west Asia. The SE trade winds of the South Indian Ocean no longer meet opposing winds when they reach equatorial regions. They cross the equator, veering to become the steady SW monsoon winds of the northern Indian Ocean and China Sea. A similar modified form of monsoon also occurs along the West African coast in the summer months when the SE trades, having crossed the Equator, arrive as SW winds along the coast between lat. 5° and 10°N.

In the Indian Ocean the SW monsoon is considerably more stormy than the NE monsoon; in the China Sea, however, the winds are in general appreciably stronger and steadier during the latter monsoon than during the former. In general, the monsoon winds when fully established tend to be somewhat stronger than the trade winds. The SW monsoon of the Indian Ocean brings copious rainfall, particularly to the windward coasts of India and south-east Asia, and is sometimes associated with poor visibility, which is not as a rule

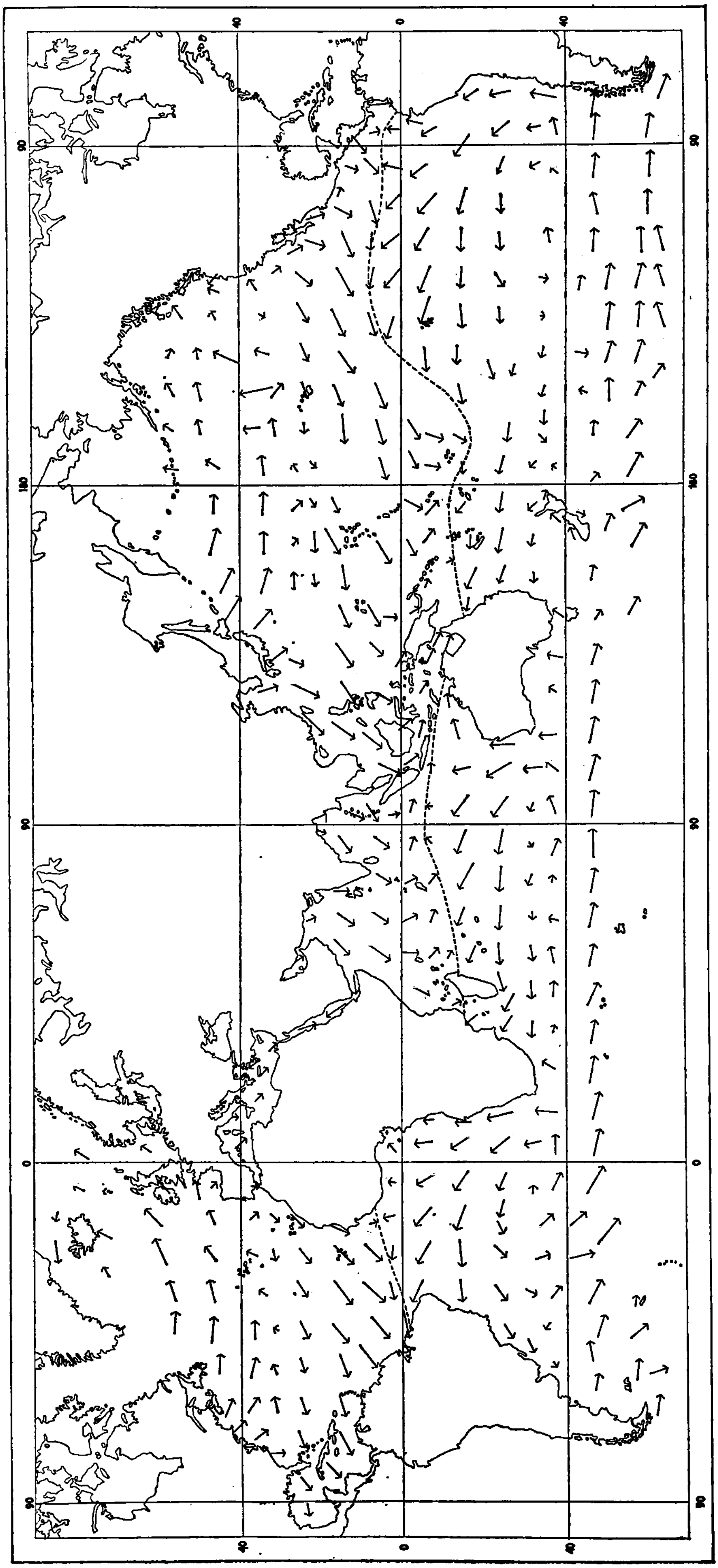

Fig. 7.3. Vector mean winds over the oceans, January

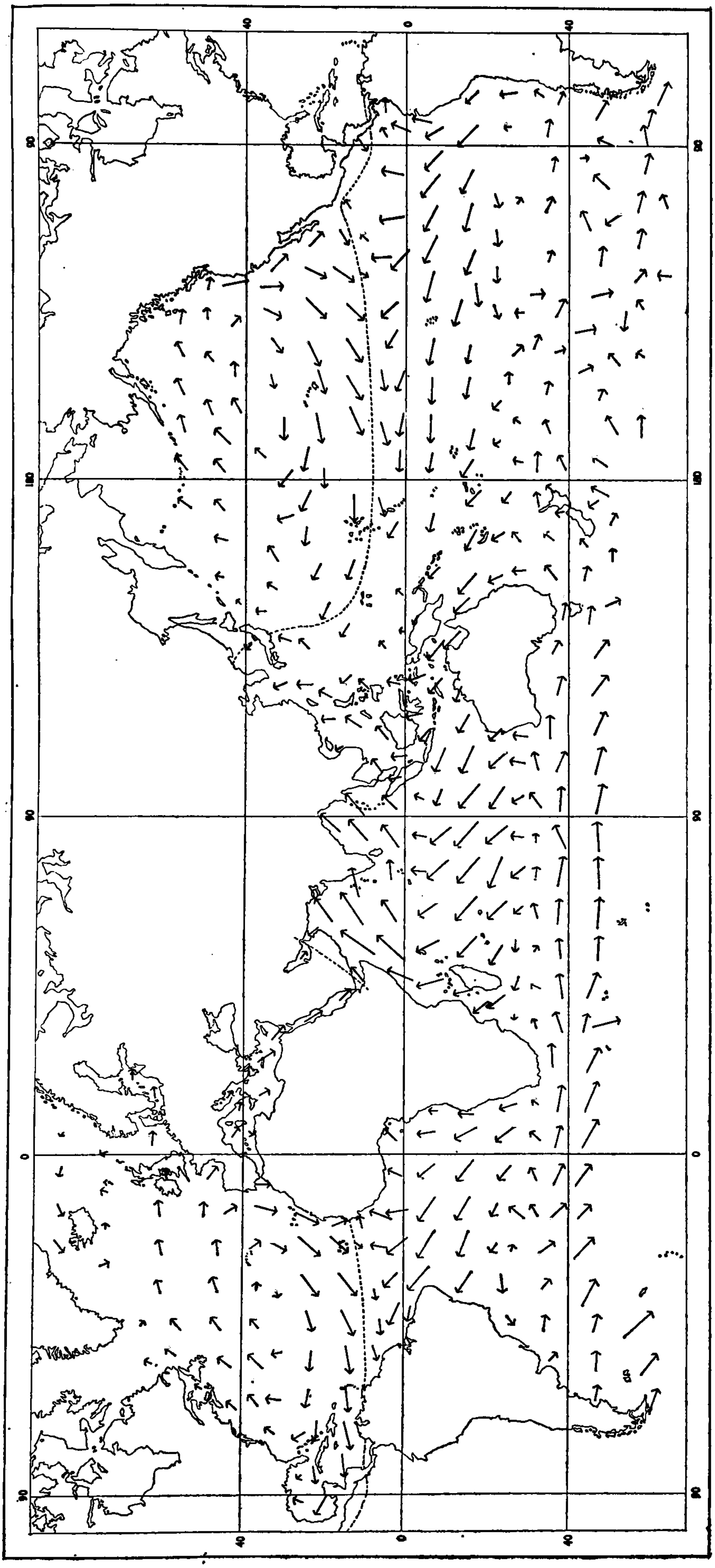

Fig. 7.4. Vector mean winds over the oceans, July

persistent. Away from the doldrums belt the NE monsoon almost invariably brings fine and clear weather. The only notable exceptions to this are along the coasts of south China and Indo-China, and less markedly along the southern portion of the Atlantic coast of North America where, in the late winter, spells of overcast drizzly weather are frequent with poor visibility. (This weather is the result of activity at the south-western ends of the Pacific and Atlantic polar fronts respectively.) Such spells may persist for up to 10 days along the south China coast in February to April, where they are locally known as Crachin and are an important climatic feature at that season.

Subtropical High Pressures

In the regions where the centres of subtropical anticyclones are most often located, that is over the middle of the oceans in lat. 30–35°N and S, there are generally calms or light winds and fine, clear weather; in fact, typical anticyclonic weather. These regions are known as the horse latitudes, from the unfortunate necessity, in sailing-ship days, of throwing overboard horses being carried to America and the West Indies, because of shortage of water and fodder, when the ship's passage was unduly prolonged due to the light winds.

Winds of the Temperate Zones

On the poleward side of the subtropical high-pressure belts are those disturbed regions where the mobile depressions and anticyclones of the temperate zone are found. As these pressure systems move, generally from a westerly direction, they cause considerable variations in wind direction and force at any given place. On the whole there is a predominance of westerly winds and a relatively high mean rate of travel of the surface air-masses in some easterly direction. In the southern hemisphere these winds so often reach gale force that they are known as the Roaring Forties, after the latitudes wherein they are generally located. Fig. 7.3 shows that in the southern summer they form a continuous belt around the globe; in winter the continuity of this belt is broken in the South Pacific where, according to Fig. 7.4, the Roaring Forties appear to be less in evidence. It is likely that this effect is real, in so far as the wind directions in this region are extremely variable and thereby the magnitude of the vector means is reduced, but it is important to remember that the information in this region is based upon scanty data.

Polar Winds

The wind systems of the polar regions have not yet been studied in detail. In north polar regions the winter winds are in most areas very variable both in force and direction; quiet periods associated with developing independent anticyclones over the polar icefields alternating with disturbed periods due to incursions of depressions formed on the Atlantic and Pacific Arctic fronts. In summer winds are generally lighter, as depressions on the Arctic fronts are less frequent and less active.

In south polar regions, winds generally have an easterly component over seas bordering the Antarctic continent. However, depressions skirting the coasts cause frequent changes of wind direction.

Katabatic winds (*see* page 89) are an important feature of local weather in many places in Arctic and Antarctic regions. They are particularly frequent and severe where high ice or snow-covered land lies close to a straight coast as in

Adelie Land in the Antarctic and in east Greenland. The former area is one of the windiest regions of the world, and the Mawson Expedition experienced a mean annual wind speed near gale force.

Land and Sea Breezes

The main wind systems of the world may be modified by local causes, of which differential heating of land and sea is one of the most important. This gives rise to land and sea breezes which, by their reaction on the general wind distribution, may modify it appreciably.

During daylight hours the land warms up much more rapidly than the sea, partly because the specific heat of soil and rock is less than that of water and partly because the sun's rays penetrate to a greater depth in water than on land, and therefore have to heat a greater mass of water than of land. Thus the air near the surface warms up more rapidly and rises more easily over land than over water. Air flowing in from seaward to replace this rising air forms the sea breeze which under favourable conditions may set in well before midday, but often does not appear till the afternoon.

At night the reverse action takes place. The air over the land cools more rapidly by radiation and, becoming heavier, flows down the slope to the surface of the sea. It displaces and forces upwards the air already over the sea, which flows landwards at a higher level to complete the circulation. The land breeze, which is impeded by inequalities in the ground, trees, houses, etc., is much weaker than the sea breeze.

Sea and land breezes, especially the former, appreciably modify the climate near the coast. The diurnal range of air temperature at the seaside is usually appreciably smaller than further inland.

Sea breezes are of some importance near coasts, where in favourable circumstances they may reinforce a gradient wind and increase the wind to above gale force. This can happen on western coasts of Europe and America in summer, and on tropical coasts when the monsoon is well developed and skies are clear. In the tropics the effect of the land breeze is often felt several miles to seaward, and the sea breeze can affect wind direction over the sea for 20 miles off the coast.

Anabatic Winds (Greek *ana* = up, *baino* = to move)

This is the name given to winds which blow up the slopes of a mountain or the sides of a plateau in calm, sunny conditions. They are found most commonly on steep sided islands and bare mountain slopes in the cloudless arid regions of the tropics and subtropics. Anabatic winds can be considered as part of the convection currents which form in the daytime over a mountain heated by sunshine. At a later stage clouds may be formed over the mountain and showers may occur.

Katabatic Winds (Greek *kata* = down, *baino* = to move)

During the night, particularly with a clear sky, heat is radiated from the surface of the earth, which cools and consequently cools the air immediately above it. Where the ground is sloping, gravitation causes this cooler, denser air to flow down the slope, forming a katabatic wind, which may have no relation to the distribution of atmospheric pressure. In mountainous countries these winds can be violent ; for example, in Greenland katabatic winds up to storm force locally sometimes blow down the slopes to the sea.

Local Winds

The following are some brief notes on important local winds.

BORA. A cold, dry, katabatic wind which blows, from directions varying between north and east, down from the mountains of the north and east shores of the Adriatic. It is often dangerous, as it may set in suddenly and without much warning, the wind coming down in violent gusts from the mountains. It may occur behind the cold front of a depression, or when there is an intensification of the winter anticyclone over the continent to the north. It occurs chiefly in winter, when it may attain gale force.

GREGALE. A strong NE wind, found in the central and western Mediterranean. It has a significance for the mariner at Malta and on the east coast of Sicily, where the chief harbours are open to the NE. It occurs mainly during the winter. At Malta it generally occurs with pressure high to the north and low to the south, but may also occur after the passage of a depression.

HARMATTAN. A dry E wind, which blows on the west coast of Africa, between Cape Verde and the Gulf of Guinea, in the dry season (November to March). Coming from the Sahara it brings clouds of fine dust and sand, which may be carried hundreds of miles out to sea. Being dry and comparatively cool, it is a pleasant change from the normal damp tropical heat of this district in spite of the dust it brings, and is known locally as the 'Doctor'.

LEVANTER. An E wind in the Strait of Gibraltar. It brings excessive moisture, cloud, haze or fog, and sometimes rain. It is usually of only light or moderate strength, and under such conditions the 'Levanter cloud' usually stretches from the summit of the Rock for a mile or so to leeward; when it is fresh or strong, violent eddies, which are troublesome to sailing craft, are formed in the lee of the Rock.

MISTRAL. The Mistral is a strong N or NW wind which blows over the Gulf of Lions and adjoining coastal districts, particularly the Rhône Valley. It usually occurs with high pressure to the north-west, over France, and low to the south-east, over the Tyrrhenian Sea, but the wind due to this pressure distribution is strengthened by the katabatic flow of air down the mountains, and by the canalizing of the air down the narrow, deep Rhône Valley. In spite of warming caused by compression during this descent, the Mistral is usually a cold wind, but it is dry, so that the weather is usually sunny and clear. The Mistral often reaches the sea as a gale and rapidly produces a rough sea.

NORTHER. The Norther of Chile is a N gale, with rain, which occurs usually in the winter, but occasionally at other times. It generally gives good warning of its approach, i.e. a falling barometer, a cloudy or overcast sky, a swell from the northward and water high in the harbours, while distant land is unusually visible.

The term Norther is also applied to strong, cool, dry N winds which blow over the Gulf of Mexico and western Caribbean, chiefly in the winter. These Northers occur when there is an intense anticyclone over western North America and one or more depressions off the eastern seaboard of the United States. The Norther frequently sets in suddenly, with no warning. These winds sometimes reach gale force in the Gulf of Mexico, but as they travel southwards into the Caribbean their strength diminishes.

PAMPERO. This is the name given, in the Rio de la Plata area, to a line-squall occurring at the passage of a sharp cold front. It is usually accompanied by rain, thunder and lightning, while the wind backs suddenly from some northerly direction to a south or south-west direction.

Pamperos are most frequent between June and September. They can generally be foretold. Pressure falls slowly, while the wind blows at first from some northerly direction, with high temperature and humidity. As the cold front approaches the winds become NE, strong and gusty. Other signs of the approach of the Pampero are said to be a rising of the water level in the river, and increase in the number of insects in the air and the extreme clearness of the atmosphere.

The wind dies away for a short time prior to the arrival of the SW squall. This is usually accompanied by large Cu or Cb cloud, torrential rain and a fall in temperature. The SW wind may be very severe, up to 70 knots or more, during the first squall, but moderates afterwards.

SCIROCCO. This name is given to any wind from a southerly direction in the Mediterranean. Since it originates from the deserts of North Africa it is hot and dry at most times of the year when it leaves the African coast. On its way north this air picks up a large amount of moisture from the sea, and is regarded as a disagreeable wind by the inhabitants of many countries of the northern Mediterranean on account of its enervating qualities. It frequently causes fog in these areas.

SHAMAL. In the Persian Gulf, the Gulf of Oman and along the Makran coast, the term Shamal denotes any NW wind, whether it is the normal prevailing wind or a gale associated with a depression. The average direction of the Shamal is NW, but this varies from place to place, according to the trend of the coast, and may be W or even SW. During a Shamal the air is generally very dry and the sky cloudless, but visibility is bad because of the masses of fine sand and dust blown from the desert. During summer months the Shamal seldom exceeds force 7, but in winter it often reaches force 8, sometimes force 9, and at this season may be accompanied by rain-squalls, thunder and lightning. The barometer does not, as a rule, give any indication of the approach of a Shamal.

SOUTHERLY BUSTER*. On the south-east coast of Australia the S wind behind the cold front in a trough of low pressure often starts with a violent squall. It is then known as a Southerly Buster. It occurs mainly in summer, but also in spring and autumn. As a rule, warning of a Buster is given only an hour or so beforehand by the appearance of Cc and Cb clouds in the south or south-west, sometimes accompanied by lightning. A long Cu roll then appears on the horizon. The wind drops to a calm and then becomes southerly, frequently blowing with gale force, accompanied by a rapid fall in temperature.

SUMATRA. Squalls from the SW known as Sumatras occur several times a month between May and October in the Malacca Straits and the west coast of Malaya, but less frequently in the vicinity of Singapore. The squalls develop over the Straits and their approach can usually be seen in advance as they are almost invariably accompanied by a thunderstorm. At the onset the wind abruptly increases in a squall with gusts sometimes well over gale force, quickly followed by torrential rain. Sumatras generally occur between late evening and soon after sunrise, and are a distinct menace to small vessels.

TORNADO. This term is applied to two wind phenomena of different natures.

In the West African area it refers to the squall which accompanies a thunderstorm. This generally occurs with heavy rain and often lasts only a short time, but may do much damage. Tornadoes of this sort usually move from east to west and occur in the transition periods between the wet and dry seasons.

* Formerly called the BRICKFIELDER.

The other type of tornado is a violent whirl of air a few hundred feet in diameter in which cyclonic winds of 200 m.p.h. or more blow near the centre. The whirl is broadest just under the cloud base and usually tapers down to where it meets the ground. In appearance it is mostly described as a dark funnel-shaped cloud like a long rope or snake in the sky, its darkness being due to the combination of thick cloud, rain, dust and other debris resulting from the destruction caused along its path. These tornadoes occur most frequently in the middle west and central plains of the United States. The chance of any one locality being struck is small as the path of a tornado is usually less than 15 miles long, and the average number over the whole United States is 140 a year. Conditions most favourable for their occurrence are when maritime-polar air from a north-westerly direction overruns maritime tropical air from the Gulf of Mexico, leading to a steep lapse rate and great instability, and then formation of tornadoes is most likely some distance ahead of the surface cold front. They can also occur in other localities, chiefly in temperate zones, about 50 having occurred in England in the past 80 years but these were all less intense than the United States variety. When a tornado moves from a land to a water surface it quickly takes on the characteristics of a waterspout, though it seldom maintains the intensity which it had over land.

Besides the tornadoes which have travelled from land to a water surface, there are numerous WATERSPOUTS* which originate over the open sea. The conditions favouring their formation over the sea are similar to those over land, namely, proximity of a frontal surface, marked changes of air and sea temperature over a short distance and above all great instability in the lower layers. Waterspouts are more frequent in the tropics than in temperate latitudes. Although they are less violent than tornadoes ashore, waterspouts are nevertheless a general hazard to all shipping and constitute a real danger to small craft.

A waterspout forms under the lower surface of a heavy Cb cloud. A funnel-shaped cloud appears, stretching downwards towards the sea. Beneath this cloud the water appears agitated and a cloud of spray forms. The funnel-shaped cloud descends till it dips into this spray cloud; it then assumes the shape of a column, stretching from sea to cloud.

The diameter of a waterspout may vary from 20 ft to 200 ft or more. The height, from the sea to the base of the cloud, may be as little as 200 ft but is usually 1,000 ft to 2,000 ft. A waterspout may last from 10 minutes to half an hour; it travels quite slowly and its upper part often travels at a different speed from its base, so that it becomes oblique or bent. It finally becomes attenuated and the column breaks at about one-third of its height from the base, after which it quickly disappears.

Frequency of Strong Winds

The frequency distribution of winds of Beaufort force 7 and higher in January and July over the oceans is shown in Figs. 7.5 and 7.6.

The greatest frequencies of strong winds occur mainly in the temperate latitudes. In the northern hemisphere the incidence of strong winds is noticeably greater in the winter months but in the southern hemisphere the difference between summer and winter is less well marked. These strong winds are associated with travelling depressions and can blow from any direction, although winds with westerly components predominate.

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* *See* also *Marine Observer's Handbook*, Chapter 11.

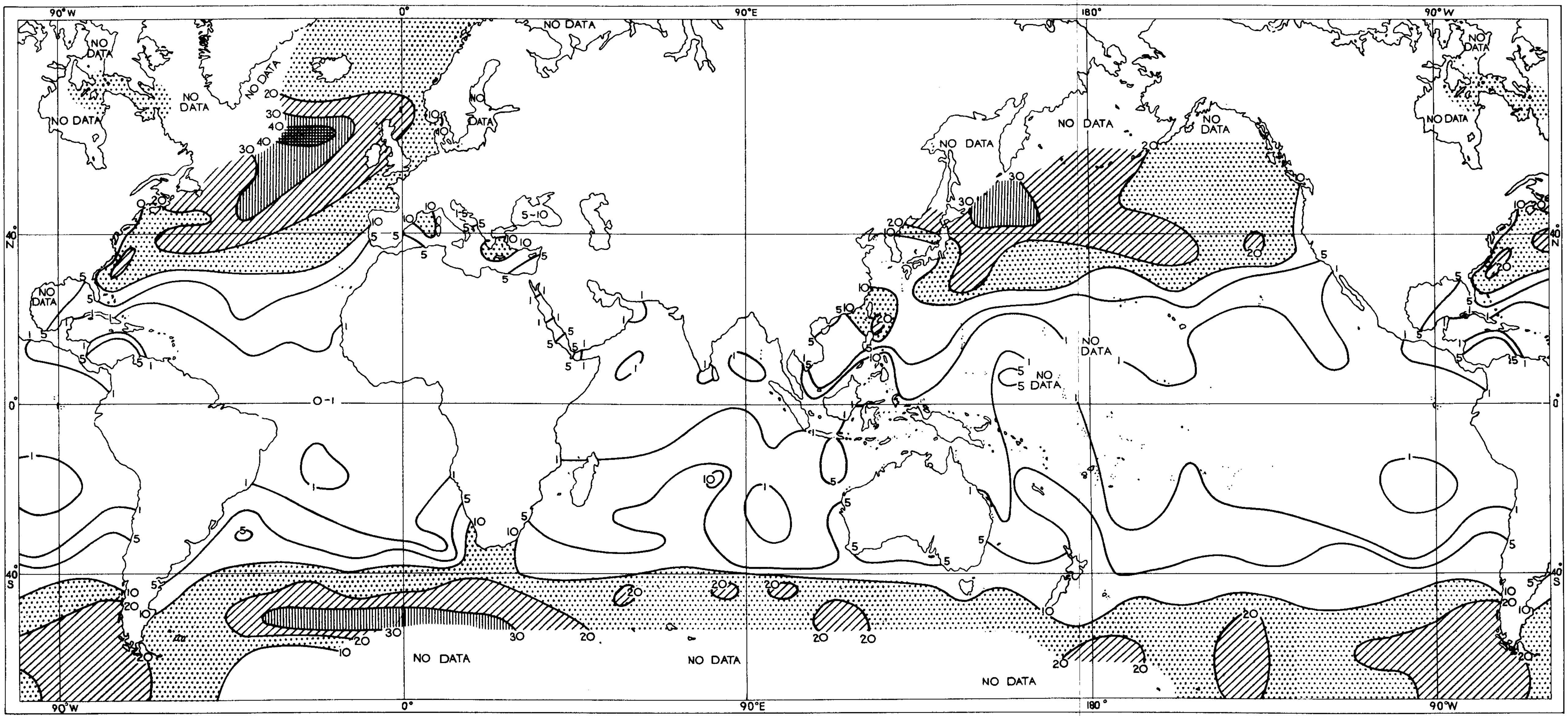

Fig. 7.5. Percentage frequency of winds of Beaufort force 7 and higher, January

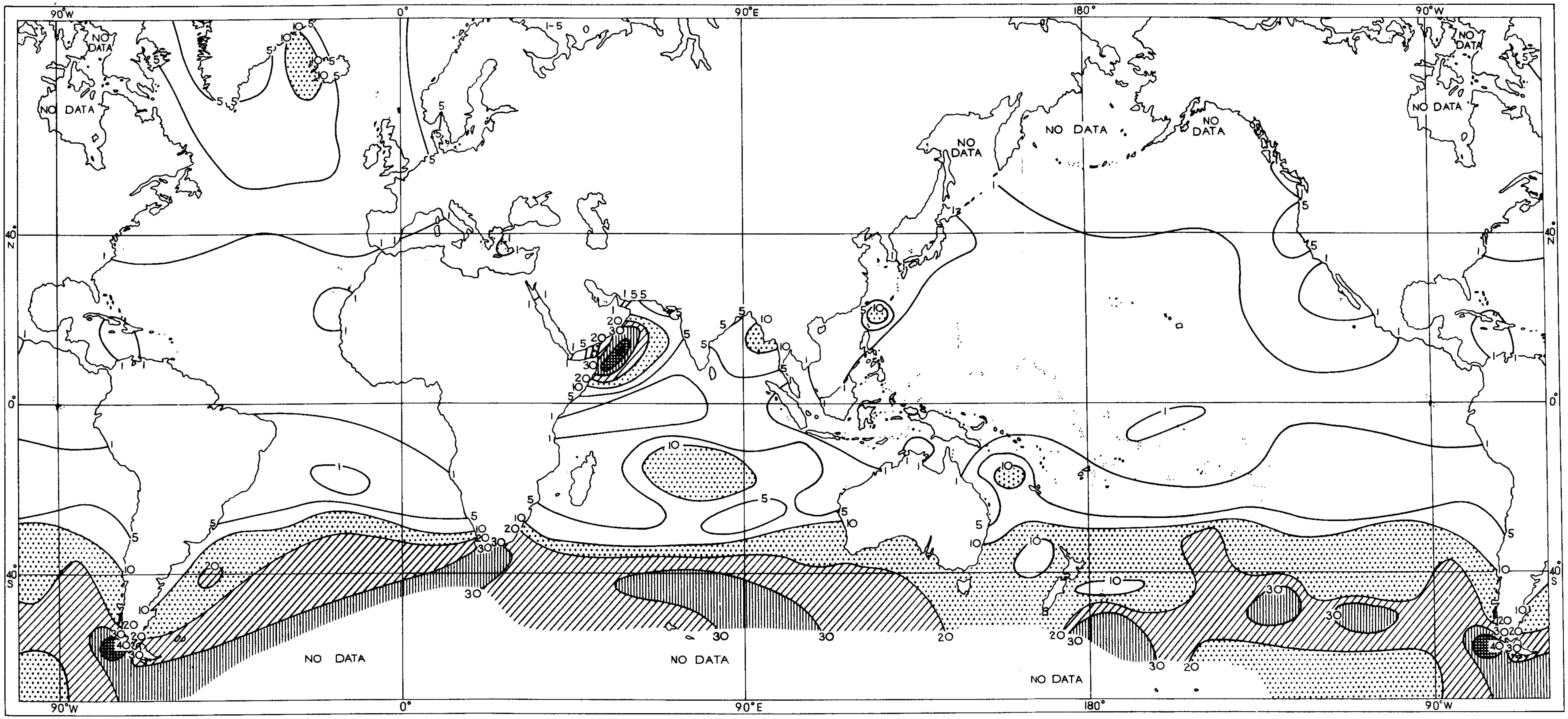

Fig. 7.6. Percentage frequency of winds of Beaufort force 7 and higher, July

Strong winds are also quite frequent in the Arabian Sea from June to August. They are associated with the south-west monsoon and, in contrast to the variable westerlies of the temperate latitudes, these strong winds are fairly constant in direction, blowing mainly from the south-west.

In general Figs. 7.5 and 7.6 show relatively low frequencies of strong winds in tropical regions but there are a few small regions showing frequencies in excess of 10%. The strong winds of these regions are associated with tropical storms, (typhoons, hurricanes, etc.) which are described in chapter 11.

Upper Air Circulation

Over the earth as a whole observations in the upper air are scanty compared with the amount of information available at the surface of both land and sea. Until the outbreak of the Second World War practically all the information available was provided by pilot balloon ascents, which seldom reached 30,000 ft and could only be used in clear weather. Since then radiosonde and radar wind-finding stations have been established in many areas adjacent to the main aviation routes of the world and aboard ocean weather ships in the Atlantic and Pacific. Enough information has thus become available to give a fairly definite picture of the mean circulation.

In the southern hemisphere the polar anticyclone at the surface is shallow and a low-pressure area is found at higher levels over the Antarctic, where easterly winds at the surface are replaced by westerlies at fairly low altitudes. The prevailing westerlies of temperate latitudes increase with altitude but show little change in mean direction. In the northern hemisphere, although the surface mean circulation is much more complicated than in the southern, the upper winds show a similar simplicity; westerly winds prevail, increasing with altitude. At about 18,000 ft the mean wind distribution shows a cyclonic circulation around a single vast region of low pressure. This low pressure is centred roughly over the Arctic Ocean and associated with it are pronounced troughs (with axes lying N to S) located over eastern North America and eastern Siberia. The strongest westerlies are found in the upper troposphere in middle latitudes in both hemispheres. The complex surface pressure systems of the northern hemisphere found on the surface synoptic charts diminish in complexity with height, and at moderate heights the charts usually show relatively simpler patterns in which troughs and ridges are commoner features than closed systems of low or high pressure.

In tropical regions well away from the equator the trade winds are quite shallow; westerlies are found from moderate heights right up to the tropopause, except in the summer when easterlies occur for a few months in the upper troposphere and are sometimes quite strong. Near the equator easterlies prevail in the upper troposphere apparently throughout the year, sometimes blowing strongly at monsoon periods.

Although radiosonde and radar wind stations are much fewer and observe less frequently than land or sea surface observing stations, nevertheless the upper air observations are now sufficiently numerous and frequent to enable synoptic charts of the upper air to be drawn as routine. Most forecasting offices use both surface and upper air synoptic charts when preparing weather forecasts. This aspect of the use of upper air information is described in Chapter 12.

CHAPTER 8

GENERAL CLIMATOLOGY OF THE OCEANS

Air and Sea Temperatures

The air temperature near the surface of the ocean is largely controlled by that of the surface waters below it. The air and surface-water isotherms at sea are therefore somewhat similar in shape.

Figs. 8.1 and 8.2 show the world's sea surface isotherms and Figs. 8.3 and 8.4 the corresponding air isotherms in January and July. To explain the distribution of these isotherms, and of certain other meteorological elements, reference needs to be made to the chart of the surface currents of the oceans facing page 202.

Except in areas such as the Red Sea and Persian Gulf, where monthly mean temperatures of the air and sea surface may occasionally exceed 90°F, the highest monthly mean temperatures in ocean areas are found, as might be expected, in tropical regions. Mean temperatures of sea or air are usually below 84°F, except in small areas where the maximum value may even reach 88°F, in tropical regions. North and south of these equatorial regions temperatures decrease, slowly at first and then more rapidly, till sea temperatures reach the freezing point of normal sea water (about 29°F) in the Arctic and Antarctic, while air temperatures in these regions may fall far below zero, particularly over icefields in winter. The small value of the extreme range of air temperature over the sea is in marked contrast to the large range experienced over the land. Examination of records up to the end of 1964 shows that the highest air temperature ever recorded was 136°F, at Azizia, Tripolitania, in 1922. This value is however very doubtful and the highest official air temperature ever recorded is 134°F at Death Valley, California, in 1931. The lowest air temperature whose accuracy is beyond reasonable doubt was – 126·9°F, which was recorded at the Soviet Antarctic base, Vostok, in 1960.

The decrease in temperature between equator and poles over both land and ocean is by no means regular. It is affected by land distribution, by winds blowing from the land, particularly as regards air temperatures, by warm and cool surface currents which occur in all oceans, and by upwelling of colder water from below the surface.

One of the most striking examples of temperature irregularities is in the North Atlantic. The Gulf Stream, moving up the American coast and then across the Atlantic as the North Atlantic Drift, causes the isotherms to run not east and west, but almost SW to NE. The cold Labrador Current, moving southwards between the Gulf Stream and the American coast, complicates the isotherm pattern still more and creates, south of Newfoundland, the steepest sea temperature gradients in the world.

In the North Pacific there is a similar phenomenon. The warm Kuro Shio Current, after flowing north-eastwards along the coast of Japan, moves eastwards across the ocean as the North Pacific Current, while the cold Kamchatka Current comes southward to meet it off Japan. However, the resulting distortion of the isotherms to east of Japan is not so marked as in the North Atlantic.

The effect of ocean currents on isotherms is also shown along the east coast of South America, where the warm Brazil Current moves south to meet the cold Falkland Current; and on the west coast where the relatively cool Peru Current flowing to the northward mixes with colder water which has upwelled from below. (*See* Chapter 16.)

Other examples of upwelling occur off the coast of South-West Africa where, for some distance off shore, the isotherms run almost parallel to the coast: off the coast of Mauritania and Senegal and, in summer, off the coast of California and Oregon.

As a general rule sea temperatures all over the world tend to be slightly higher than the overlying air temperature. The exceptions are chiefly in the western North Atlantic and western North Pacific in summer, in regions like the Red Sea and Persian Gulf in summer where air temperatures tend to become abnormally high, and in regions of upwelling, e.g. the coasts of Morocco, South-West Africa and South America.

Precipitation over the Oceans

Although much experimental work has been done, so far it has not been possible to make accurate measurements of rainfall at sea. The occurrence of precipitation (rain, snow, hail) has, however, been regularly recorded aboard observing ships, and charts have been drawn showing the number of times precipitation of all types has been recorded in a given area, expressed as a percentage of the total number of observations. Charts for January and July are shown as Figs. 8.5 and 8.6.

In the Atlantic there is an area of high rainfall frequency in the doldrums, caused by the convergence of the NE and SE trades. This is centred about long. 25°W and has a movement north, between January and July, of about 5° in latitude. North and south of this area the precipitation frequency decreases, especially off the coast of Africa; the frequency is also quite low in the Gulf of Guinea, well within the tropics. As we pass from the regions of the subtropical highs to those of the mobile depressions of the temperate zone, the frequency increases again, particularly in winter.

In the eastern Pacific there is an area of high rain frequency in the doldrums, centred about 125°W. Unlike the corresponding Atlantic area, its seasonal movement in latitude is very small. The general distribution of precipitation is otherwise similar to that of the Atlantic.

In the western Pacific the distribution is more irregular. It is complicated by orographic rains due to the islands of the East Indies and Philippines and by the change, in the China Sea and neighbouring regions, from the NE monsoon of winter to the SW monsoon of summer.

In the Indian Ocean the January chart may be taken as typical of the NE monsoon months, December to March, and even of April. A large area of relatively high precipitation frequency near the equator is caused by the interaction of the SE trades with the north-westerly continuation of the NE monsoon across the equator. Precipitation falls off considerably in the Arabian Sea and Bay of Bengal and off the African coast south of 8°N. To the south it diminishes in the subtropical high-pressure belt of the southern hemisphere, particularly off the north-west coast of Australia and to the west of South-west Africa, but increases again in the regions of the temperate zone depressions.

As the SW monsoon sets in, rainfall frequency increases considerably on the windward coasts of India and Burma, which show values up to 50% in July. The rainfall there is mainly orographic; leeward coasts or sheltered regions have generally little rain. Most of the tropical and subtropical South Indian Ocean shows a fairly high frequency, except near Australia and Africa, while further south we find the usual winter increase of precipitation in the temperate and sub-Antarctic zones.

Mean Cloud Amount over the Oceans

As might be expected, the isopleths of mean cloud amount over the oceans follow much the same pattern as those of rainfall (Figs. 8.7 and 8.8). Where the mean cloud amount is high with rainfall frequency low, as in the South Atlantic west of Angola (West Africa), this is probably due to an unusually large amount of cloud of the stratus types.

In the Atlantic the doldrums region is cloudy. The amount decreases polewards, especially near the coasts, except off Angola. In higher latitudes cloud increases again, particularly in winter.

In the eastern Pacific the distribution is similar. There is also a region off the South American coast, similar geographically to the Atlantic region off Angola, where the mean cloud increases while the rainfall frequency decreases.

In the Indian Ocean the distribution of cloud amount is broadly in accordance with that of rainfall frequency, both in January and July.

Distribution of Sea Fog

Fog is comparatively rare in the tropical and subtropical regions of the oceans, since conditions favourable for the advection of air sufficiently warm and moist relative to the temperature of the water surface do not exist in these regions. The distribution of fog over the oceans in January and July is shown in Figs. 8.9 and 8.10, from which it can be seen that the frequency of fog may reach 40% in certain localities. Some of the regions of relatively high fog frequency at sea are:

(*a*) THE NEWFOUNDLAND AREA. The reason for the high frequency of fog off the Great Bank of Newfoundland is explained thus: the cold Labrador Current flowing southward through the Davis Straits meets the warm Gulf Stream as the latter begins its journey across the Atlantic. When the winds in summer are from a southerly direction, they become heavily loaded with moisture as they pass over the warm waters of the Gulf Stream. They are chilled by this cold Labrador Current and the moisture they contain is condensed to form fog. If the wind shifts to the west or north-west the fog quickly clears. In the winter the winds are more often westerly or north-westerly; coming from the land, and having only a short journey over the sea, they are comparatively dry, and the fog frequency is consequently less at that season.

(*b*) THE NORTH-WEST PACIFIC. There is a similar explanation for the high frequency of fog in the north-west Pacific. The warm Kuro Shio, on its way north-eastwards across the Pacific, is met by the cold Kamchatka Current. The winds in summer are southerly and therefore warm and charged with moisture; cooled by the cold Kamchatka Current they cause frequent fogs. In winter the winds tend to be

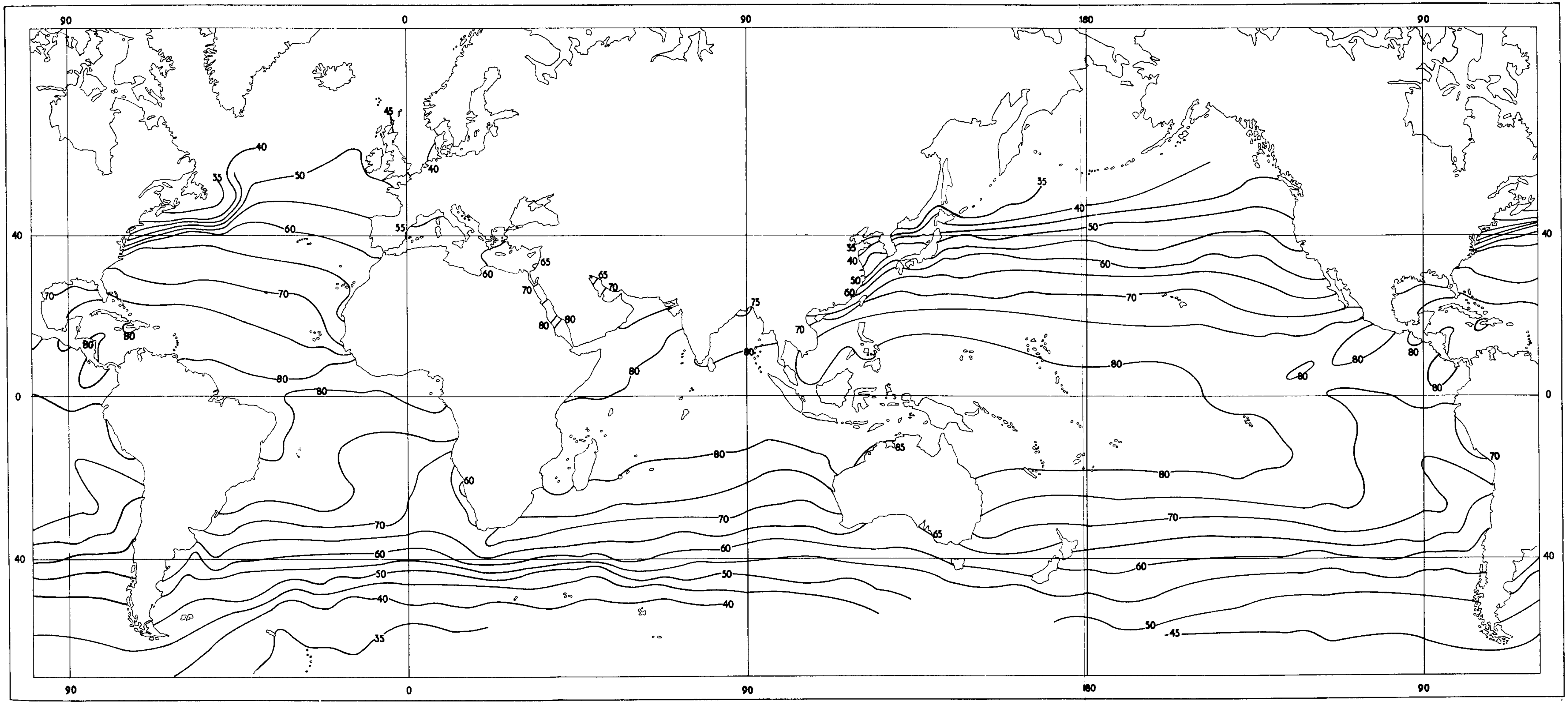

Fig. 8.1. Mean sea-surface isotherms in °F, January

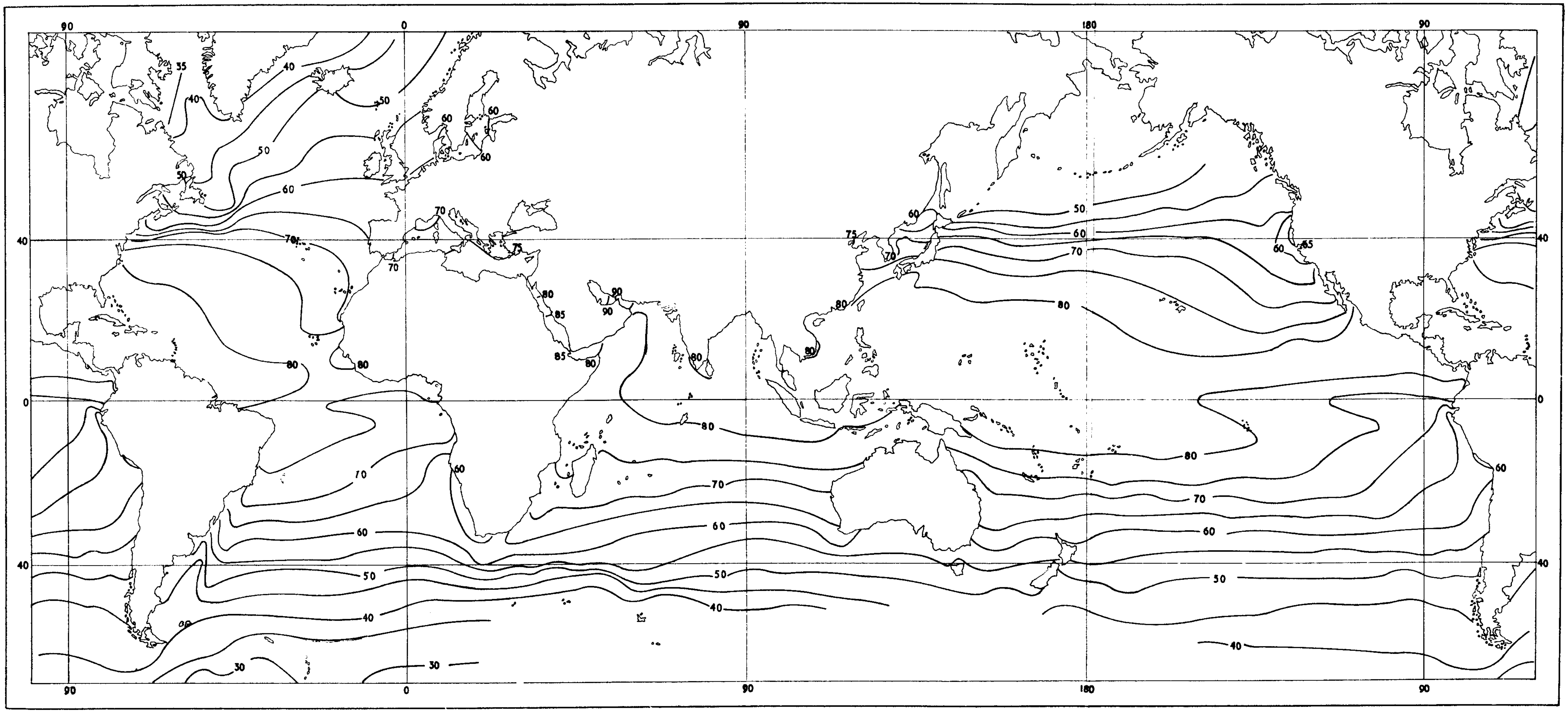

Fig. 8.2. Mean sea-surface isotherms in °F, July

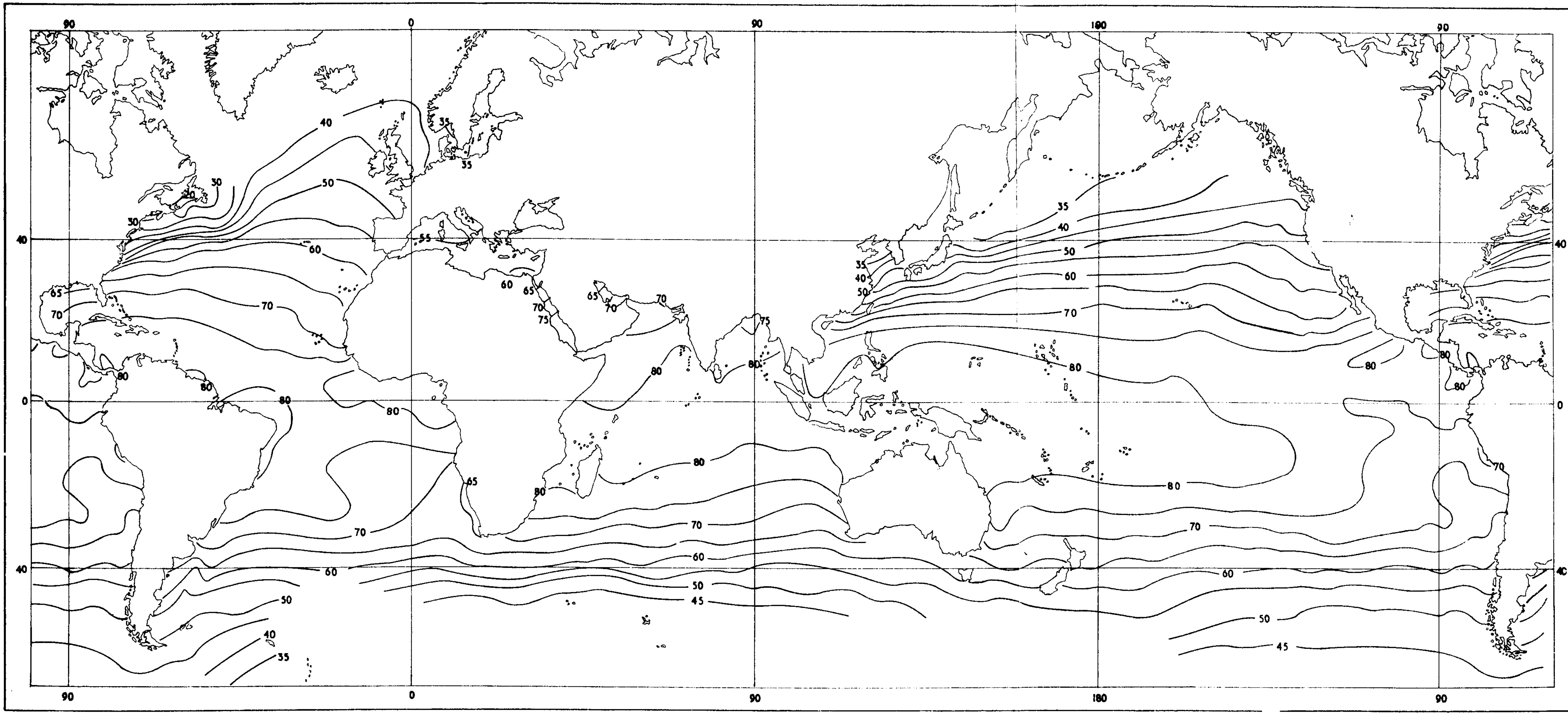

Fig. 8.3. Mean air isotherms in °F over the oceans, January

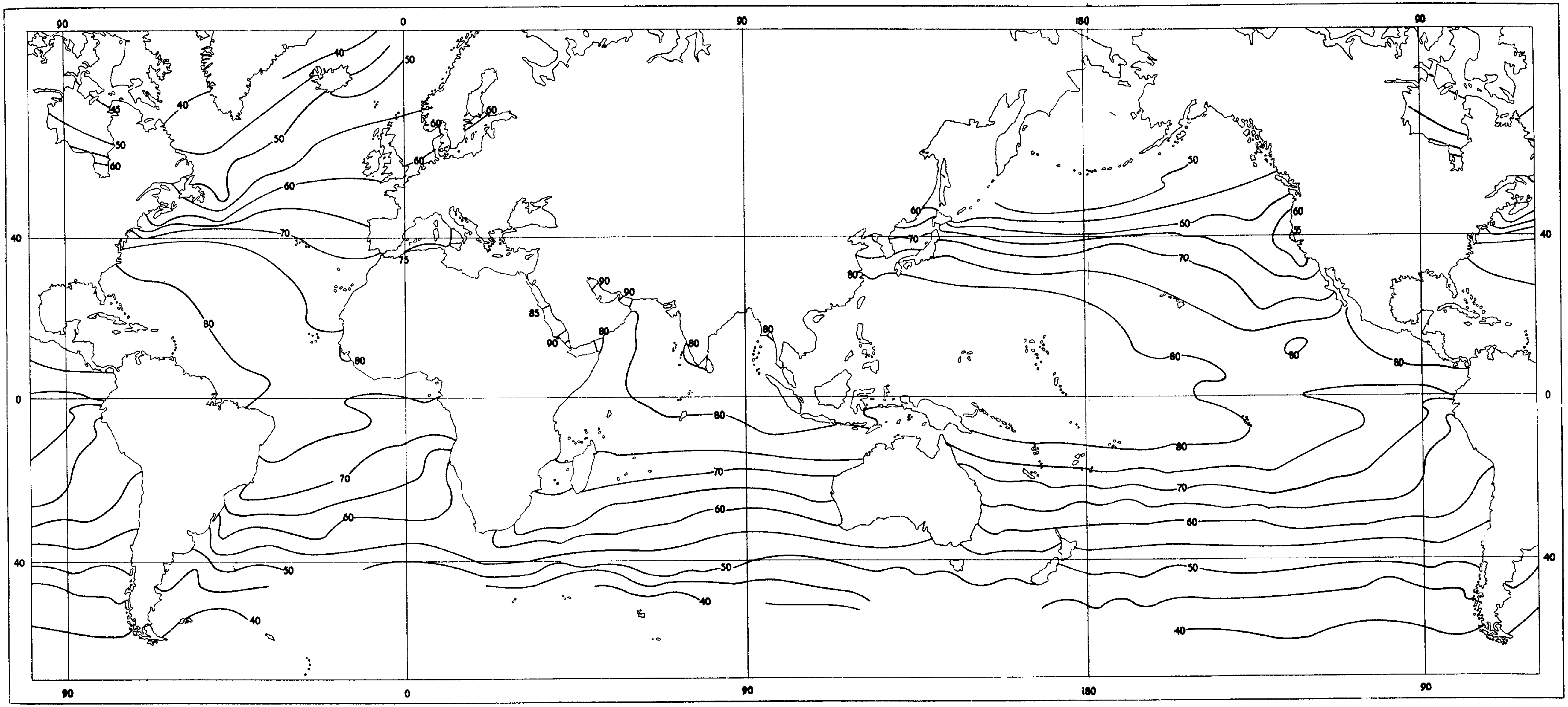

Fig. 8.4. Mean air isotherms in °F over the oceans, July

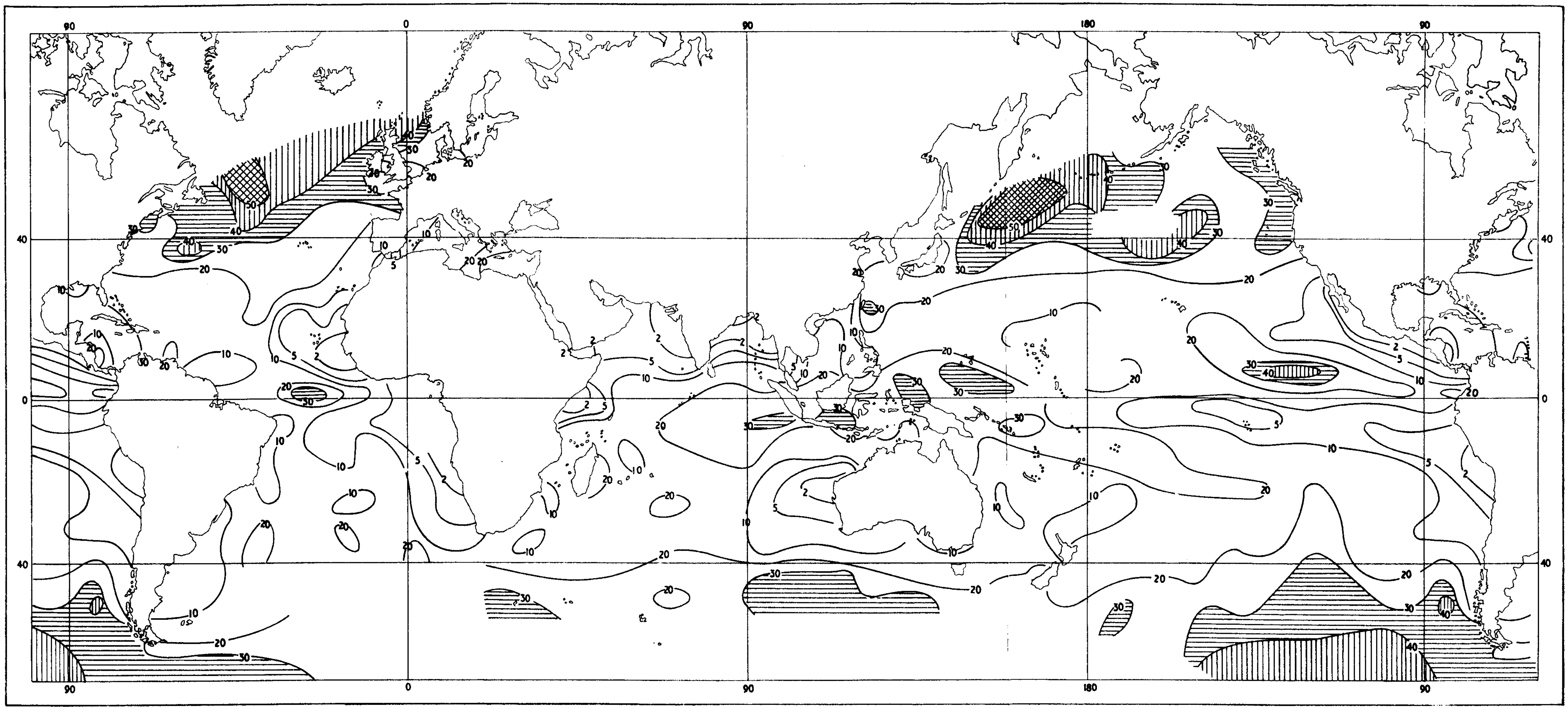

Fig. 8.5. Percentage frequency of precipitation over the oceans, January

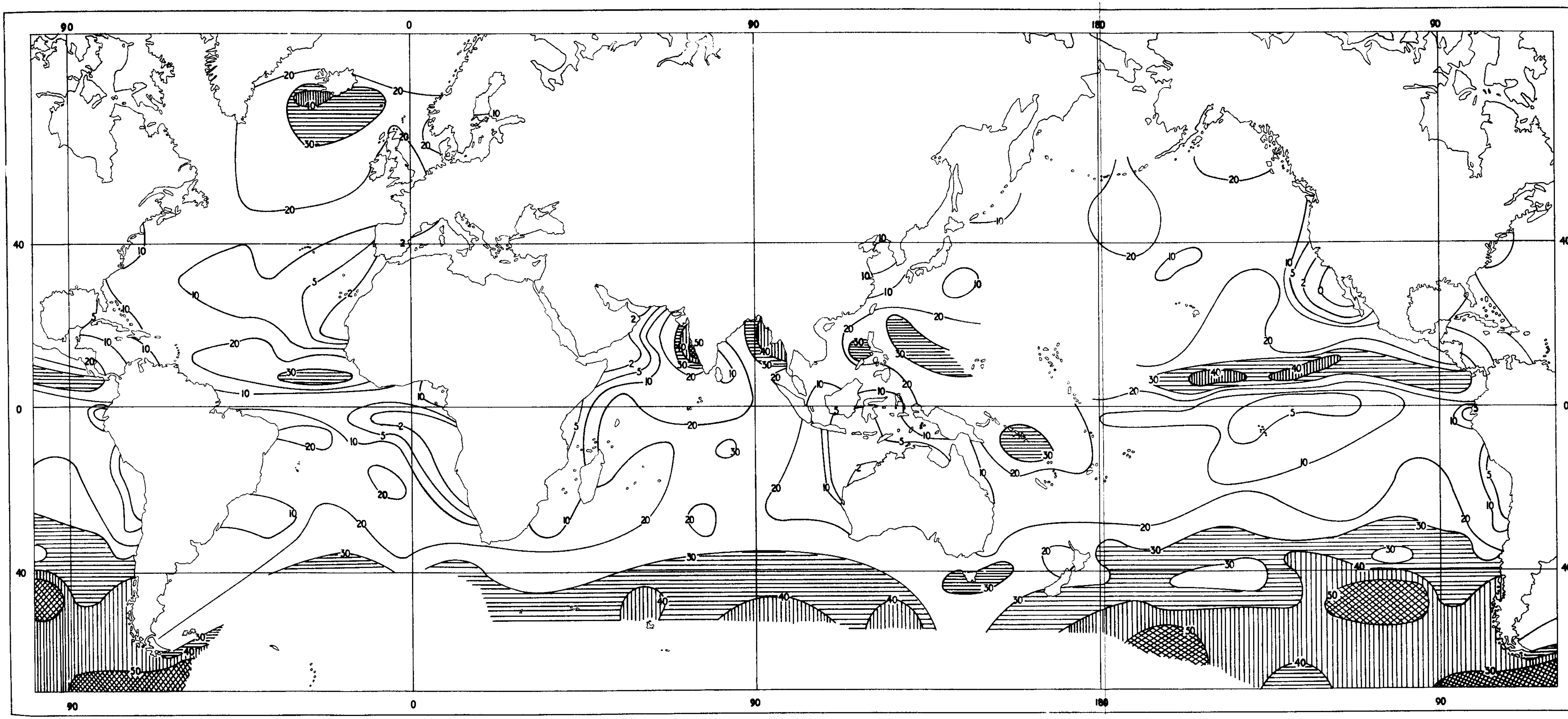

Fig. 8.6. Percentage frequency of precipitation over the oceans, July

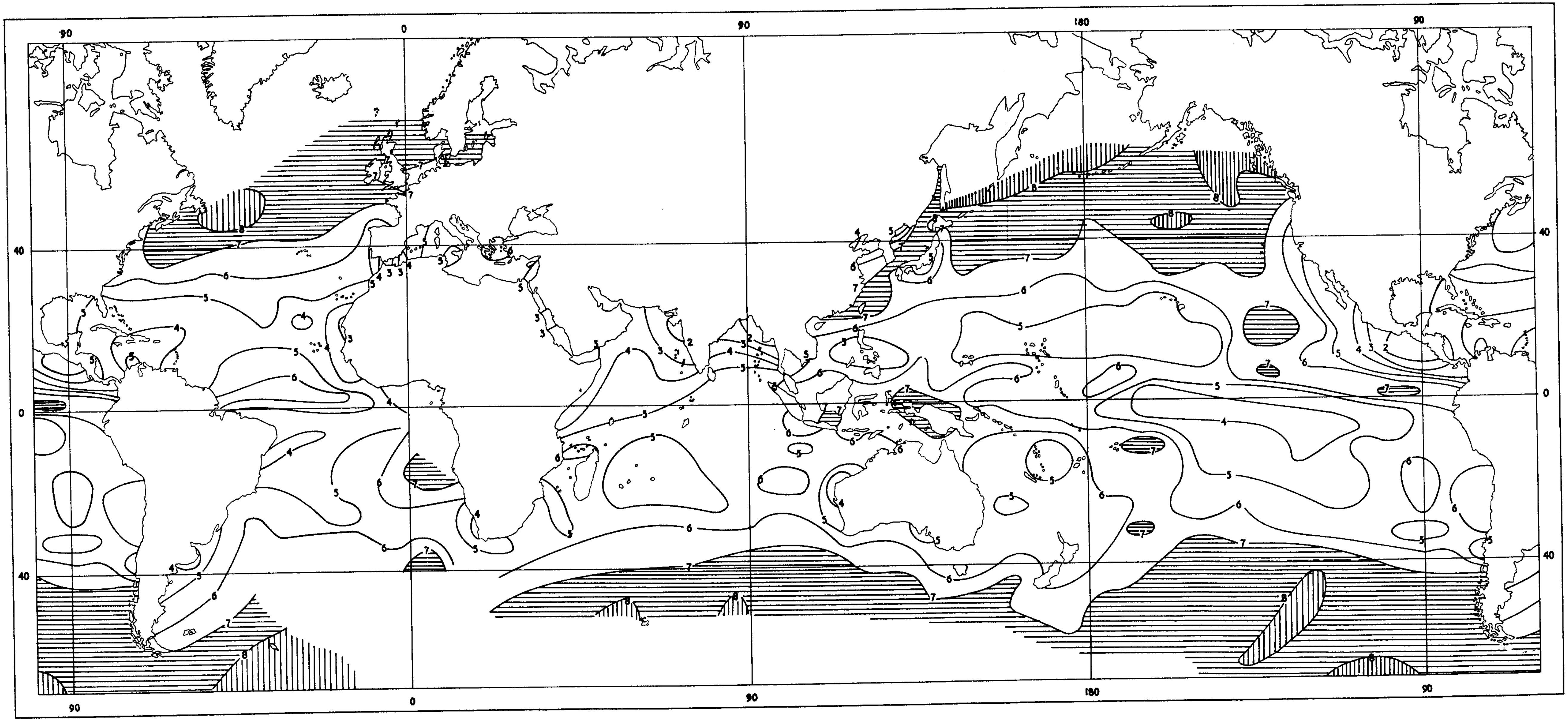

Fig. 8.7. Mean cloud amount in tenths over the oceans, January

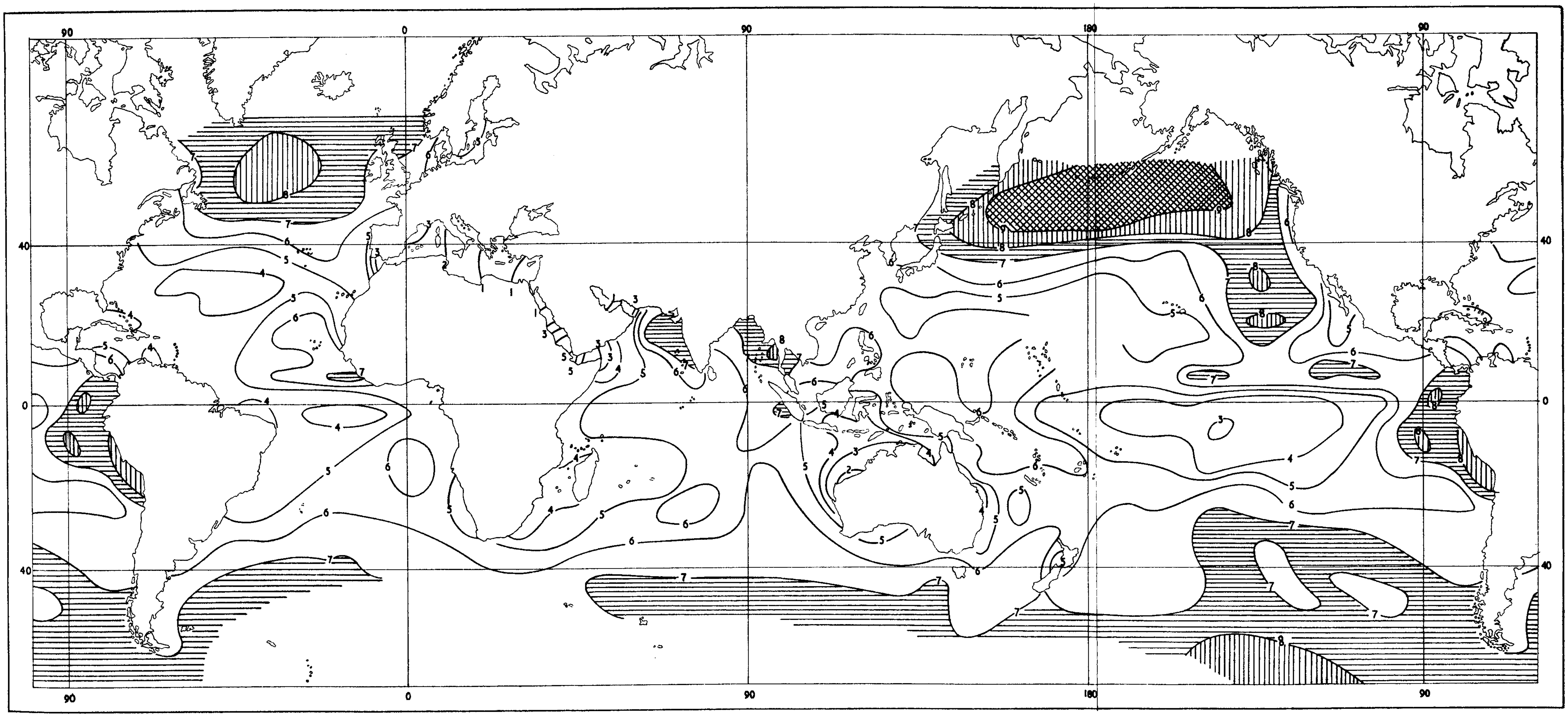

Fig. 8.8. Mean cloud amount in tenths over the oceans, July

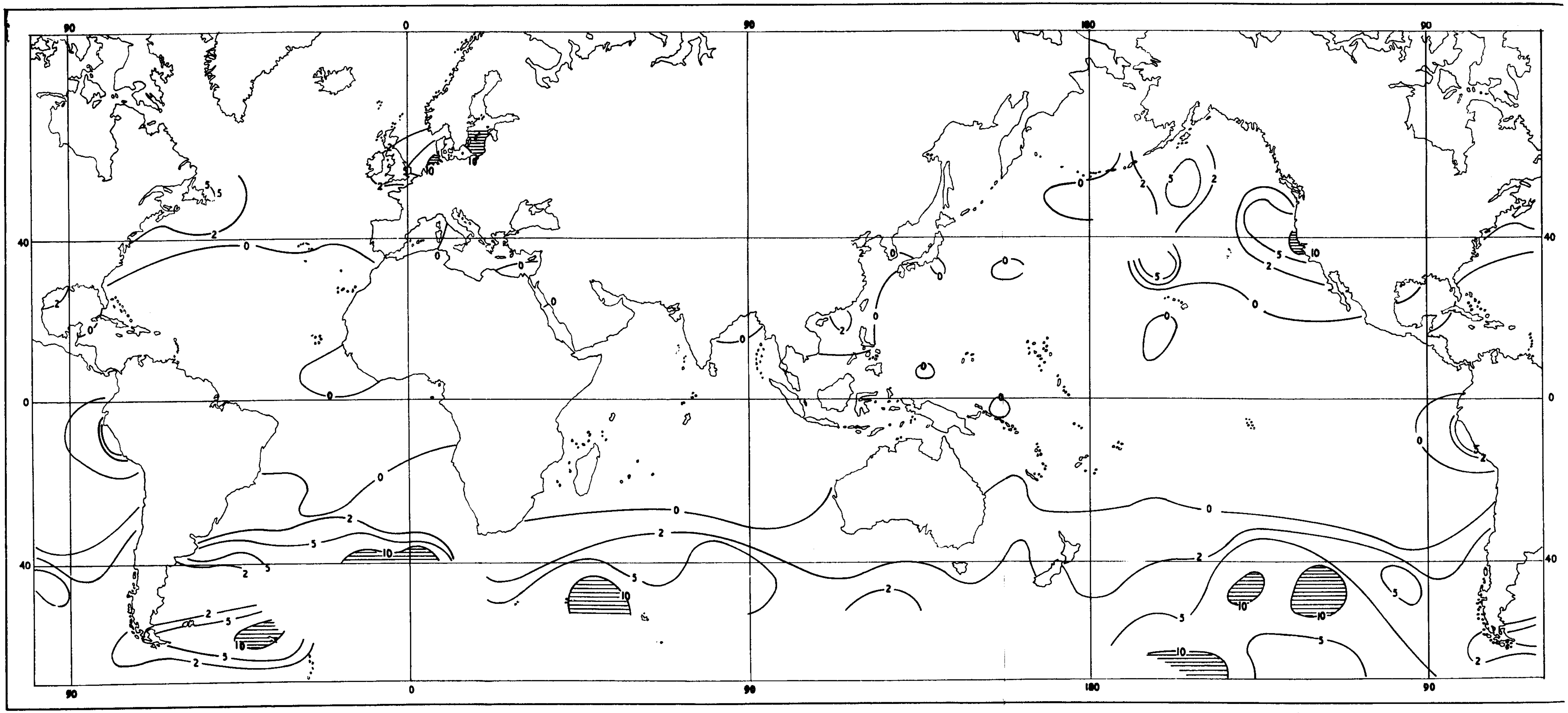

Fig. 8.9. Percentage frequency of fog over the oceans, January

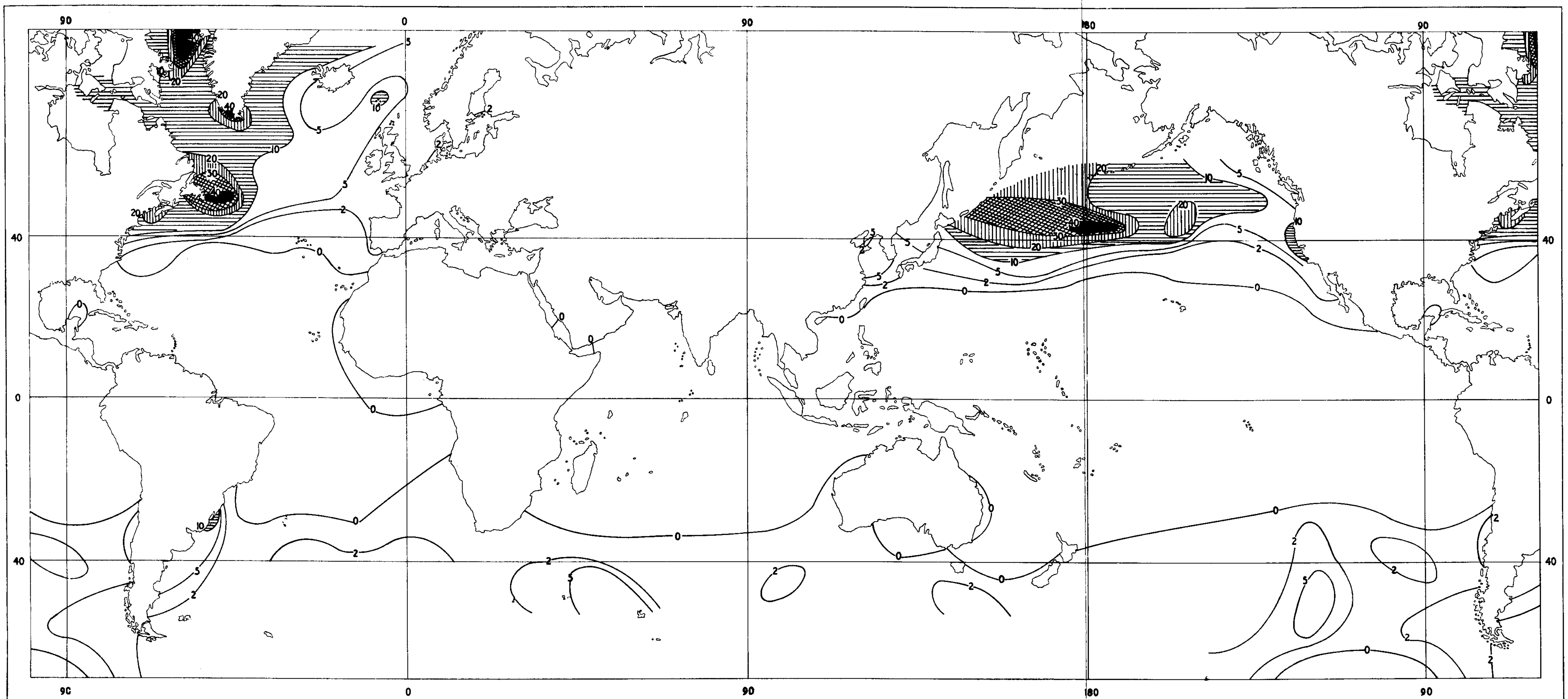

Fig. 8.10. Percentage frequency of fog over the oceans, July

more westerly or north-westerly and therefore cool and dry, with only a short sea journey, and fogs are generally less frequent at that season.

(*c*) The Subtropical West Coasts of the Continents. These include the coasts of California, Morocco, Chile and South-west Africa. Here the relative coldness of the water depends on upwelling from beneath the surface, due to the removal of surface water near the coast by the trade winds, which blow parallel to the shore. The Californian coast is the most foggy of these regions throughout the year, and the frequency of fogs there is worst in the summer and early autumn. The highest frequency occurs close to the coast and does not show up well in Figs. 8.9 and 8.10.

(*d*) The Polar Regions in Summer. In summer many parts of the Arctic regions are covered by areas of relatively low pressure. As a result, southerly winds frequently bring large amounts of warm moist air into the region, which is soon chilled to the point where fog results. In particular, the regions where pack-ice persists throughout the summer are frequently affected by fog.

Similar areas of appreciable fog frequency occur over the southern oceans in their summer due to southerly advection of relatively warm moist air over the cold sea, particularly near the edges of the pack-ice. Along the coasts of the Antarctic, however, fog is less frequent than over the sub-Antarctic regions of the southern oceans. Katabatic winds blowing from the interior of the continent are common along the coasts even in summer. Fog is most common in the pack-ice and when warmer air penetrates to the sub-Antarctic regions, as often happens near occluded fronts which have been carried around the southern side of dying depressions in these regions. Falling snow anywhere in the polar regions is also often responsible for reducing visibility to an extent which interferes with navigation.

PART III. WEATHER SYSTEMS

CHAPTER 9

STRUCTURE OF DEPRESSIONS

Introduction

In earlier chapters where the physical properties of the atmosphere and its general circulation have been described, the emphasis has chiefly been laid upon wind and pressure distribution. In the remaining chapters dealing with meteorology we shall be more concerned with the dynamical structure and physical characteristics of pressure systems such as the depression and the anticyclone, as the reader needs this preliminary knowledge before he can properly understand the principles of weather forecasting which are outlined in the later chapters.

We have already referred to depressions and anticyclones as the main features on a weather map, and have shown that they are associated with the movement of very large amounts of air between low and high latitudes, as well as with large-scale transfers of air at higher levels more nearly parallel to the equator, and mainly from west to east. As will be shown later, it is reasonable to assume that the behaviour of pressure systems is, to some extent, related to the physical characteristics of the air currents entering into and partaking in their circulation; and whenever these air currents have originated from widely separated latitudes, it is also largely dependent upon the *difference* between the physical characteristics of these air currents. Before a more detailed study of depressions and anticyclones is made, closer attention must be paid to these large-scale movements of air.

Air Masses

A study of air movement on a large scale would be simplified if the air flowing over an area of several thousand square miles could be given a broadly descriptive title in accordance with one or more of its physical properties, assuming for this purpose that these properties had a nearly uniform value over the whole area. To be of any use, such a classification would have to be based on temperature and water vapour content, since these are the physical properties of air of most concern to the meteorologist. At first sight the reader may well find it hard to believe that any properties of the atmosphere could be really uniform over such a wide area. He could reasonably point to the objection that temperature varies rapidly with height, and that the water vapour content of the atmosphere decreases rapidly with height.

However, the temperature and water vapour content of air usually change much more slowly in a horizontal direction than in a vertical. In fact study of synoptic charts has shown that sometimes air can acquire and retain substantially the same values of temperature and water vapour content at any one level over areas comprising thousands of square miles. In any such area, the temperature and moisture characteristics of one sample air column is very much like that of any other, since the horizontal homogeneity at the surface largely imposes a corresponding homogeneity at each level aloft. The whole amount of air overlying such a region is termed an AIR MASS, though the use of the term is usually restricted to air within the troposphere. The next steps are to see how air masses are formed and to look for the regions where their formation takes place.

Source-regions

The essential characteristic of an air mass is that the distribution of temperature and humidity is broadly uniform throughout the air mass in a horizontal plane. Air masses are formed therefore over regions where the earth's surface temperature is nearly uniform and the winds are comparatively light. These factors ensure that air can remain in the region long enough to acquire the characteristic physical properties, which are largely determined by the nature of the underlying surface. For instance, the oceanic areas usually covered by the central regions of the subtropical anticyclones favour the formation of uniform air masses. Areas which produce this effect on the air above them are known as SOURCE-REGIONS. Other typical source regions are snow-covered continents, the Arctic Ocean and extensive deserts such as the Sahara.

Figs. 9.1 and 9.2 show very broadly the locations of the principal source-regions of air masses for limited areas in January and July respectively. The mean flow patterns of the air masses from these source-regions are indicated in a general way by continuous lines and arrows; the air masses thus indicated are of course those that occur *most frequently* in the areas concerned. The study of synoptic charts shows that the patterns of flow of these air masses on a given day will show marked differences from Figs. 9.1 and 9.2, and that many areas will often be occupied by air masses different from those thus indicated. The largest variations in the positions of air masses are found in the temperate latitudes in both northern and southern hemispheres. Some indication of this variability is given for certain regions in Figs. 9.1 and 9.2, where the paths most frequently followed by the other air masses which sometimes reach these regions have been shown by dotted lines and arrows.

Classification of Air Masses

Air masses are classified as follows:

(*a*) An absolute classification based on the principal source-regions, in which the following descriptive terms are used to describe the air masses:
  - (i) Arctic (A)
  - (ii) Maritime polar (mP)
  - (iii) Continental polar (cP)
  - (iv) Maritime tropical (mT)
  - (v) Continental tropical (cT)
  - (vi) Equatorial (E).

(*b*) A relative classification based on the influence sustained, or modifications undergone, by the air masses when they leave the typical source-regions.
  - (i) COLD AIR MASSES, whose surface temperature is below that of the underlying surface.
  - (ii) WARM AIR MASSES, whose surface temperature is above that of the underlying surface.

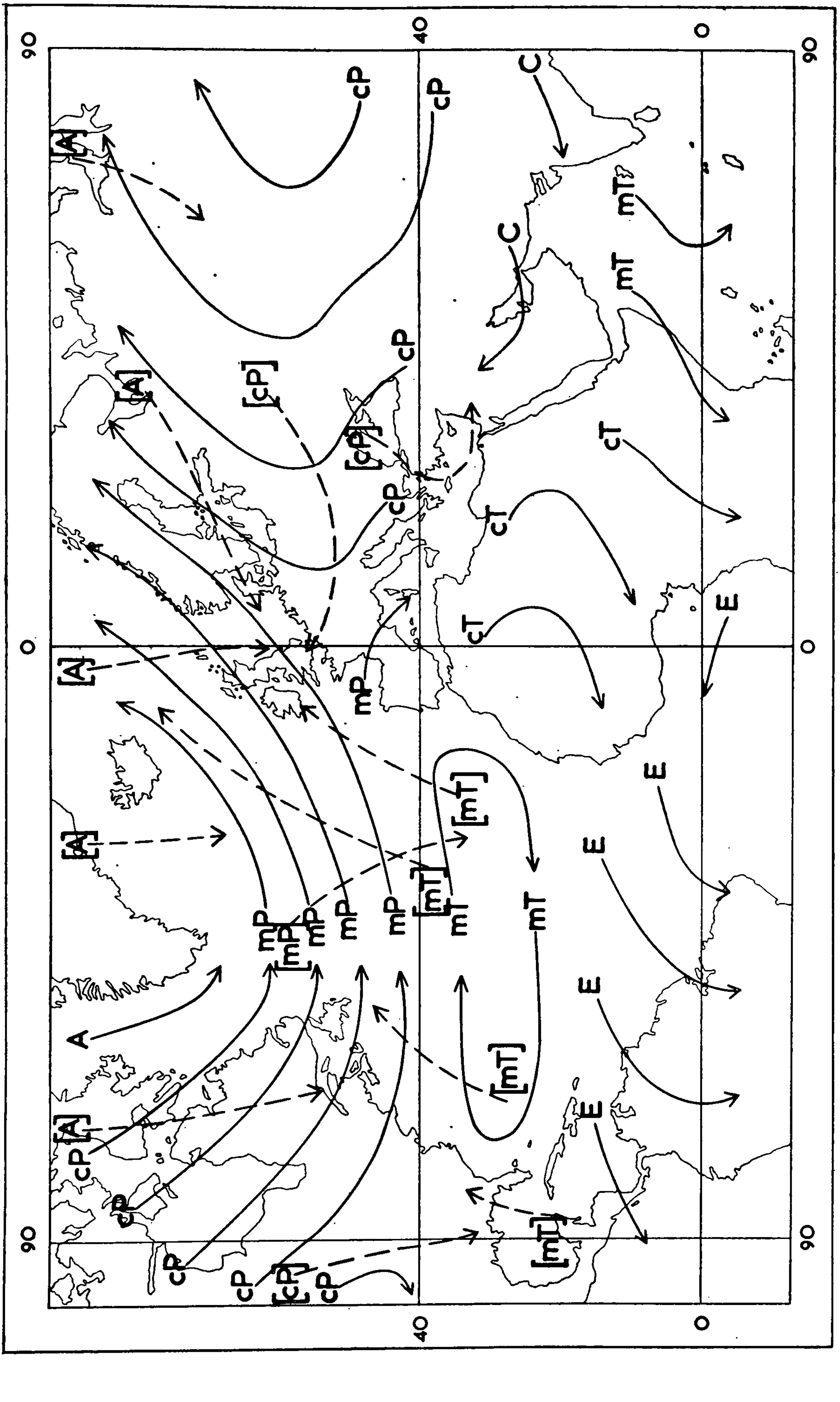

Fig. 9.1. Principal air-mass source-regions, January

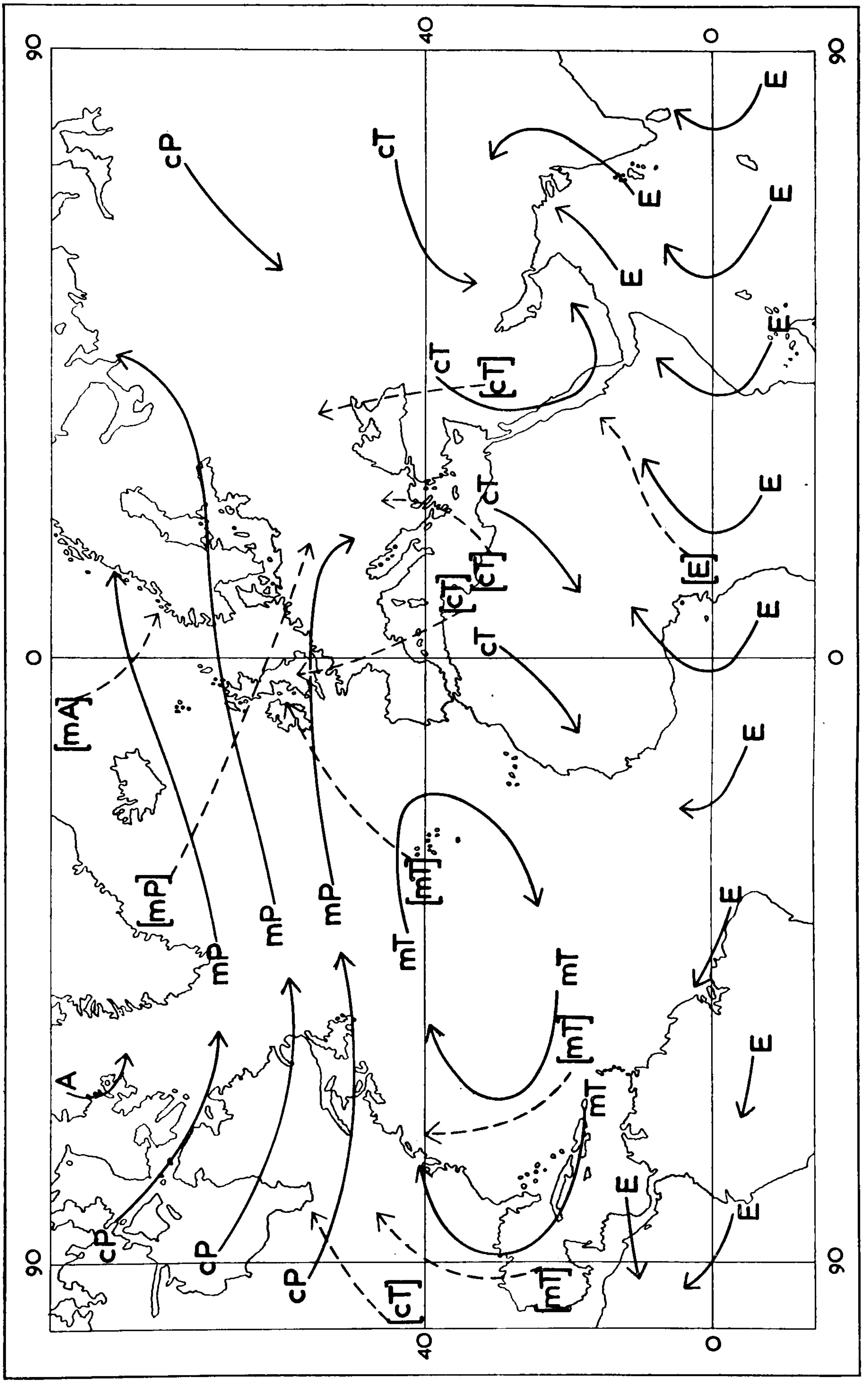

Fig. 9.2. Principal air-mass source-regions, July

Thus a cold air mass is one that is being heated from below, and a warm air mass one that is being cooled from below. In general an air mass is labelled according to the source-region which it last occupied.

Life History of Air Masses

As soon as it leaves a source-region the properties of an air mass begin to be modified. These modifications mainly result from changes in the nature of the underlying surface, and changes in the radiation processes to which the air mass is subjected such as the length of the day and the mean solar elevation. The effect of these influences depends upon the time which has elapsed since the air mass left the source-region; this time is known as the 'age' of the air mass.

The thermodynamical processes which produce modifications in the air mass include heating and cooling from below, and the addition or removal of water vapour by condensation or evaporation. Dynamical processes which produce modifications are convection, turbulence and subsidence.

Air-mass Properties

The properties of an air mass, while still in its source-region, may be deduced from surface conditions. Evidently ARCTIC and CONTINENTAL-POLAR AIR MASSES are cold at all seasons and extremely cold in winter, since they originate over surfaces of land or sea which are covered with extensive ice or snowfields for a large part of the year. The prevailing low temperatures result in an air mass, which originates from a large land area such as Greenland or Antarctica, having at first a very low absolute humidity, before it has had time to pick up moisture from the sea surface. Since ice and snow surfaces are good radiators of heat the maximum cooling takes place in the lowest layers of the air. In such an air mass the lapse rate in the lower layers is thus small, and often shows a marked temperature inversion close to the ground.

In winter MARITIME-POLAR AIR MASSES are usually warmer than continental-polar air in the surface layers. Thus in maritime-polar air there is often a steep lapse rate in the lowest layers, the humidity also showing a rapid decrease with altitude since the moisture is mostly added to the surface layers from the sea.

Both CONTINENTAL-TROPICAL and MARITIME-TROPICAL source-regions are warm so the overlying air is heated, though to a much smaller extent in the air of maritime-tropical origin. In maritime-tropical air masses both relative and absolute humidity are high; however, in continental-tropical air in which very high dry-bulb temperatures are common, the relative humidity is generally less than in the maritime variety.

The main modification which an air mass undergoes while it travels is due to heating or cooling from below, that is from the underlying surface. Heating from below creates a steep lapse rate of temperature, leading in turn to instability, convection and increased turbulence. The development of cumulus or cumulonimbus cloud, and showers, will follow when sufficient moisture is available in the air mass, or is added to the air mass by evaporation from the underlying surface. Strong convection generally results in good visibility, except in precipitation. Cooling from below is most effective in the layers nearest the surface and extends upwards only slowly due to the agency of turbulence, while in these circumstances convection is entirely suppressed. This is because an inversion of temperature is soon produced in the lowest layers of an air mass undergoing cooling from below. In very light winds, with little turbulence, the surface cooling of the air may be enough to cause condensation and fog. Stronger winds may create enough turbulence to prevent fog, but in such cases stratus clouds often form just below the upper limit of the temperature inversion, Surface visibility remains poor on these occasions because the particles which constitute haze are not dispersed through a great depth of air but are confined within the lowest layer where they originated.

When an air mass subsides its temperature is raised at the dry-adiabatic lapse rate. The relative humidity of the air mass also falls, partly due to adiabatic heating and partly because air is brought down from higher levels which has a lower moisture content than the air which it replaces. On the other hand, when an air mass is lifted, as in crossing a mountain range, its temperature falls adiabatically, often enough to produce condensation followed by cloud and precipitation.

Air-mass Boundaries—Fronts

Although each air mass is fairly homogeneous in a horizontal direction, the boundary zones of different air masses may be quite sharp. This is because mixing of air across the boundary between two distinct air masses having different temperature and humidity takes place rather slowly. On a weather map, the boundary zones of the different air masses are represented as lines, known as FRONTS. Thus a front may be regarded as a line at the earth's surface dividing two air masses. Boundaries between air masses are really 'zones of transition', but the small scale of the chart enables them to be regarded as lines. If we think in three dimensions, we can visualise two air masses separated by a surface whose intersection with the horizontal is the front as represented on the chart. This surface of separation is known as a FRONTAL SURFACE.

The name front was introduced during the First World War by analogy with the battlefronts. The analogy goes further, for most weather disturbances originate at fronts and the general picture, as successive disturbances move along the frontal zone, is one of 'war' between two or more conflicting air masses.

The principal frontal zones in the northern hemisphere are:

(*a*) The ARCTIC FRONT, in the Atlantic, separating Arctic air from maritime-polar air of the North Atlantic.

(*b*) The POLAR FRONT, in the Atlantic, which separates either continental-polar air of North America from maritime-tropical air of the North Atlantic, or maritime-polar air of the North Atlantic from maritime-tropical air of the North Atlantic.

(*c*) and (*d*) Similar Arctic and Polar fronts in the Pacific.

(*e*) The MEDITERRANEAN FRONT, separating the cold air over Europe in winter from the warm air over North Africa.

(*f*) The INTERTROPICAL CONVERGENCE ZONE. This zone was formerly described as the intertropical front, but the term has fallen into disuse, since the zone, which is really the region where the NE and SE trades either meet at an appreciable angle to each other, or else flow in nearly parallel currents, has little in common with the fronts of temperate latitudes, and the term Intertropical Convergence Zone has now been generally adopted in its place. (*See* Chapter 7.)

When the Intertropical Convergence Zone (ITCZ) is situated 5° of latitude or more from the equator, the trade-wind air between the ITCZ and the equator has crossed the equator and usually undergone deflection due to the rotation of the earth, so that it has acquired a westerly component. (For instance, when the ITCZ is in the northern hemisphere the SE trades cross the equator and approach the ITCZ as a SW wind, and in this case they meet the NE trades at the ITCZ which then becomes a zone of pronounced convergence.) When the ITCZ is less than $5°$ of latitude from the equator the trades of both hemispheres usually flow nearly parallel to each other, without marked convergence.

In agreement with these ideas the intensity of the weather phenomena experienced at the ITCZ can vary widely along its length. Where there is pronounced convergence at the zone (e.g. NE winds meeting SW winds as when the ITCZ is located a considerable distance north of the equator; or SE winds meeting NW winds when the ITCZ has similarly moved a considerable distance into the southern hemisphere in a particular locality), then this convergence is associated with a broad belt of thick, lofty convection cloud systems (cumulonimbus), together with numerous thunderstorms and heavy rain. On the other hand, when the two trade-wind systems flow side by side with little convergence, the meteorologist will often find it hard to identify the ITCZ on his synoptic chart; and all that the mariner may observe of the ITCZ as his voyage takes him across it may be an increase of cumuliform cloud along with some scattered altocumulus and extensive patches of cirrus, with perhaps a few scattered showers.

Again, in these circumstances a length of some hundreds of miles of the ITCZ may be replaced for a time by a broad belt of very light and variable winds; the mariner usually applies the description DOLDRUMS to regions characterized in this way. After a short time the air mass in such a region is so modified by convection and precipitation as to become moist throughout its whole depth. Such an air mass is then described as EQUATORIAL air to distinguish it from the maritime-tropical air of the trade-wind air streams. On a synoptic chart a region occupied by such an air mass is described as a 'doldrum area', and, as already stated, replaces for a time a portion of the ITCZ.

Figs. 9.3 and 9.4 show the mean positions of these frontal zones in January and July.

These frontal positions are very generalized. On any particular day positions vary to some extent from those shown on the maps, the greatest range in position and movement from day to day occurring in temperate latitudes. The latitudinal movement of air masses and fronts closely depends upon the distribution of pressure at any time. For instance, in certain synoptic situations over the North Atlantic it is not uncommon for maritime-tropical air to reach the Norwegian Sea area, nor for maritime-polar air to penetrate south of the Azores and Madeira. Similarly continental-polar air from the Siberian anticyclone occasionally reaches Hong Kong just within the tropics, where it has been known to bring air temperatures down to freezing-point. Incursions of maritime-polar air to lower latitudes similarly bring cold spells to the interiors of South Africa and South America in winter. Even more extreme conditions occur in the prairie lands of western and north-western Canada, which are largely shielded by the Rockies from the moderating influence of the maritime air masses of the Pacific. At the same time these great plains are readily invaded by continental-arctic air in winter and by continental-tropical air in summer; so the range of temperature is very great over the year, and sometimes great extremes are recorded within a period of a few days. Very similar conditions

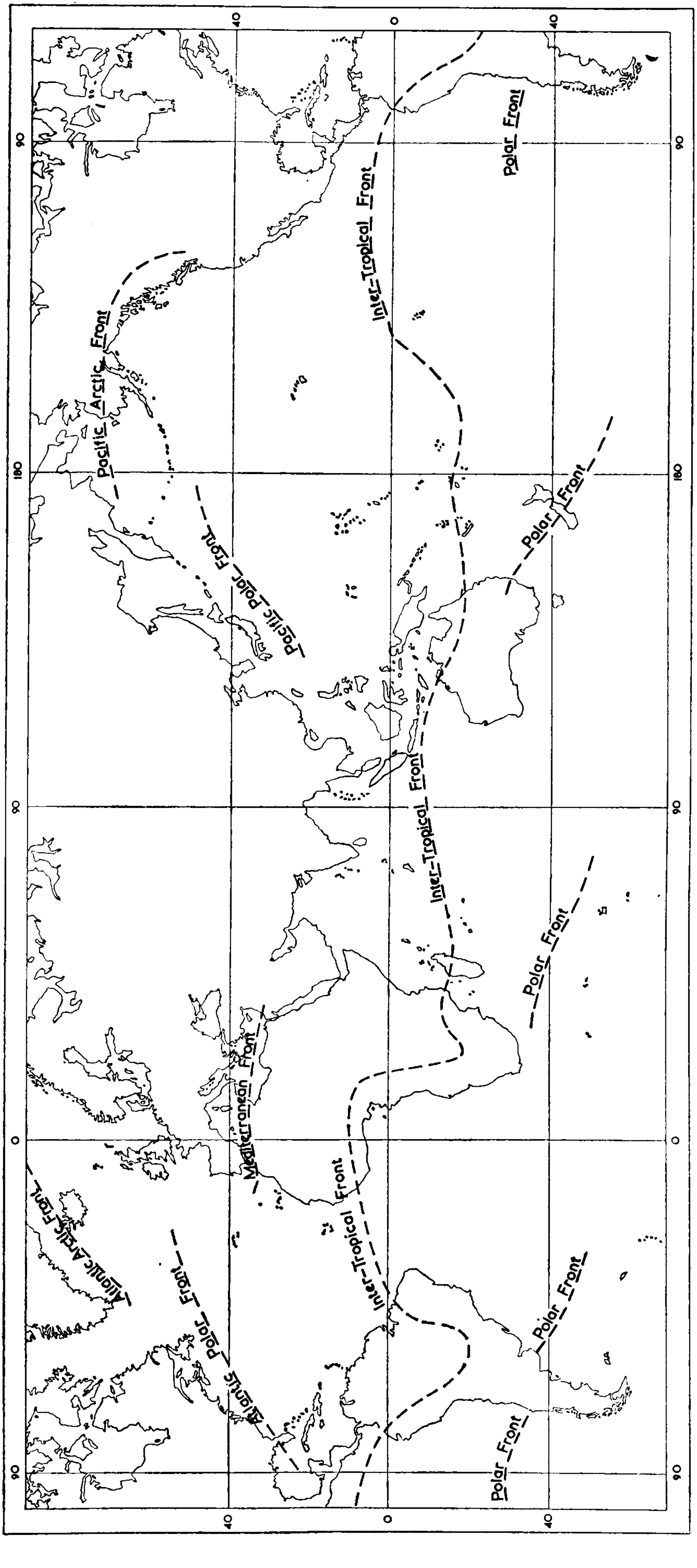

Fig. 9·3· Mean positions of frontal zones, January

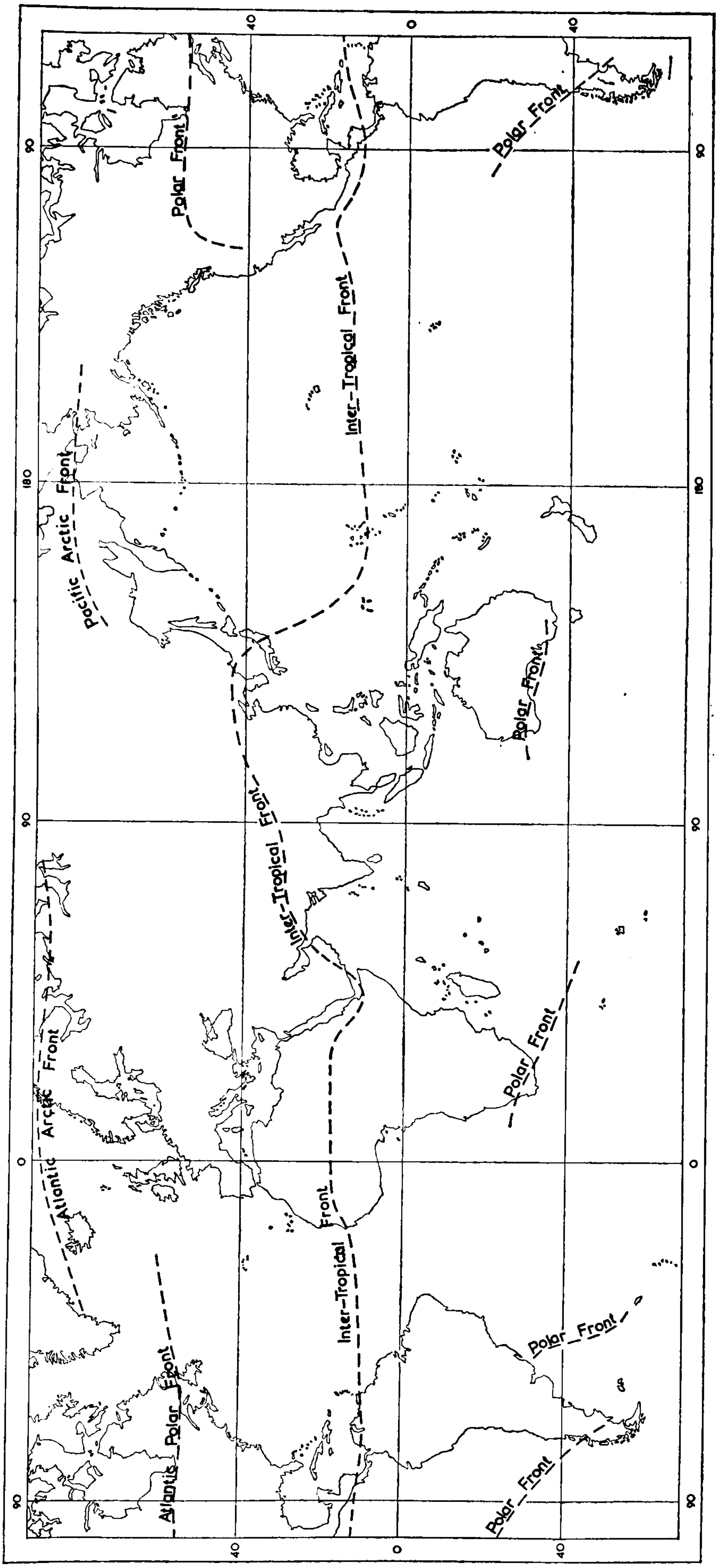

Fig. 9.4. Mean positions of frontal zones, July

occur in northern and north-eastern Siberia, though there the sheer size of the land mass and its proximity to the frozen Arctic Ocean are largely responsible for the very low temperatures experienced in winter. Other instances of the significance of air masses in local climatology could readily be cited.

Frontal Surface as a Surface of Equilibrium

A frontal surface has been defined as the surface of separation between two air masses, but nothing has been said about the nature of this surface. At first sight it is difficult to see that such a surface will exist, and the question 'Why doesn't the air mix?' arises. In fact, although the air masses do mix to some extent through their surface of separation, the scale on which this mixing takes place is so small in comparison with the magnitude of a system such as a depression, and the air masses involved in its circulation, as to make it appear as though the air flow near the frontal surfaces took place on both sides of an impenetrable boundary. In this respect the behaviour of air masses and that of the frontal surfaces separating them bears a close resemblance to that of two fluids, such as oil and water, which do not mix.

This analogy with a fluid can be extended further. For instance, it can readily be shown that any two fluids, of different density and having a relative motion, can achieve equilibrium on a rotating earth along a plane surface of separation inclined at a very small angle to the horizon. In the case of two air masses of different densities the colder, and therefore denser, air is found to lie as a wedge beneath the warmer, lighter air. The angle of inclination α is very small, its tangent being between $\frac{1}{200}$ and $\frac{1}{50}$. In Fig. 9.5, which is in section and applies to the northern hemisphere, the colder air is imagined as blowing into the paper relative to the warmer air. **FS** then represents the frontal surface and **F** the line in which this surface intersects the horizontal plane **WC**. The fact that **FS** is not horizontal is due to the rotation of the earth. The actual magnitude of α depends also on the relative velocity and difference in density of the two types of air.

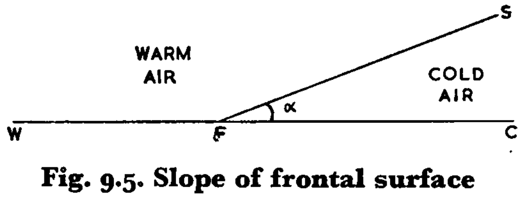

Fig. 9.5. Slope of frontal surface

Fronts occur in sea water as well as in air. One of the finest examples, with which most mariners are familiar, is the 'Cold Wall' off the Newfoundland Banks, where the cold water of the Labrador Current meets the much warmer waters of the Gulf Stream. Here the front is very sharp, so sharp in places that ships athwart it have registered sea temperatures taken from forward and aft which differed appreciably.

Convergence and Divergence

An understanding of convergence and divergence is necessary in dealing with many problems in synoptic meteorology, especially those connected with pressure change. Consider a horizontal rectangular area **ABCD** (Fig. 9.6*a*) and the space contained in an imaginary box (say 500 ft deep) erected directly over this area. The box will have vertical sides whose edges touch the ground along **AB**, **BC**, **CD**, **DA**. In Fig. 9.6*a* the arrows represent the average wind in direction and speed over the lowest 500 ft of the atmosphere. More air crosses through the vertical side above **AD** than leaves through the vertical side above **BC**, yet no air crosses the boundaries **AB** and **CD** because the wind is always blowing parallel to these lines. Air must therefore accumulate in the box unless it can escape through the top, i.e. by means of upward vertical motion. In such a case **ABCD** is described as an AREA OF HORIZONTAL CONVERGENCE and the arrangement of winds shown in Fig. 9.6*a* is an example of a CONVERGENT WIND FLOW.

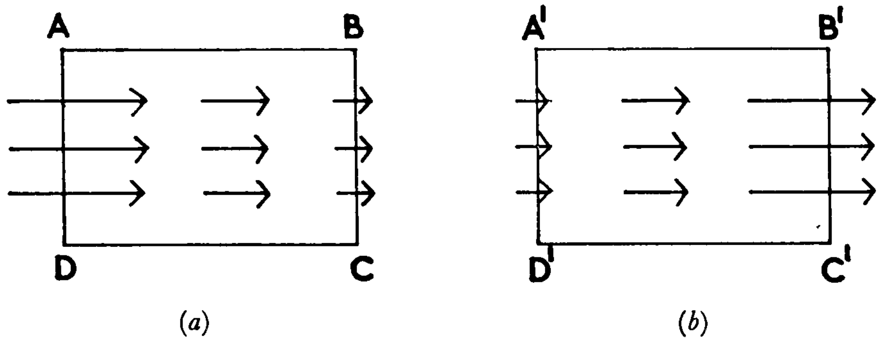

Fig. 9.6. Convergence and divergence
Length of arrow represents speed of wind

There are two points to note. A convergent wind flow does not necessarily mean that the winds all blow towards a centre; the example is an extreme case for illustration only. The essential idea is that when the horizontal flow into and out of the restricted region is computed, an excess of inflow over outflow denotes horizontal convergence. The second important fact is that horizontal convergence and upward motion occur together. Air cannot go on accumulating, and when all the horizontal motion has been taken into account, the only escape for the air is in a vertically upward direction.

Fig. 9.6*b* illustrates an example of depletion of air, in other words, HORIZONTAL DIVERGENCE. It is easily seen that horizontal divergence at the surface accompanies a downward vertical motion. This is the same as saying that the air is undergoing subsidence as described earlier in this chapter.

We have, therefore, the following results applying to motion of air near the earth's surface:

Upward motion of air is associated with areas of horizontal convergence in the lower atmosphere.

Downward motion of air is associated with areas of horizontal divergence in the lower atmosphere.

Life History of a Depression

In Chapter 2 a depression was defined as a region of relatively low pressure with closed isobars.

Shortly after the First World War, Norwegian meteorologists, headed by Bjerknes, developed a theory of the formation, growth and decay of depressions in middle latitudes. More precisely, the achievement of the Norwegian meteorologists was to gather together the results of different workers and give a coherent account of the life history of each depression. The main advantage of the theory was that it related the life of depressions to the frontal zones and air masses separating them, and thus enabled the future behaviour of a depression to be forecast from a study of its previous history, as shown on a sequence of synoptic charts. Initially, meteorologists only recognized one frontal zone, that classified earlier as 'the polar front in the Atlantic', and as a result the name POLAR FRONT THEORY was given to the Norwegian work at an early stage. It has now been generally accepted among meteorologists and is outlined in the following paragraphs.

Cold surface air of polar origin has already been shown to be separated from the warmer air of lower latitudes along a surface of separation whose intersection with the earth's surface is known as a frontal zone, or more generally as a polar front. This polar front is not continuous around the earth but is extensive enough to justify the simple description given above. It is on this polar front that depressions of temperate latitudes form. (*See* also pages 106 and 107.)

The formation of a depression is assisted by a large temperature difference between the warm and cold air masses. It is for this reason, among others, that the regions where depressions most frequently form are in the western North Atlantic and western North Pacific and in the Southern Ocean, where the horizontal gradients of air and sea temperatures are greatest, particularly in winter.

The warmer air which enters the circulation of developing depressions is supplied from the semi-permanent subtropical anticyclones located over the oceans in lat. 30° to 35° N and S. Reference to a map of the mean pressure distribution over the world will show that the Icelandic low pressures and the Aleutian low pressures are supplied respectively in this way with warm air from the North Atlantic and North Pacific anticyclones. Similarly warm air from the South Atlantic and South Pacific subtropical anticyclones feeds the southern hemisphere depressions of the Roaring Forties. (*See* Figs. 7.1 and 7.2.)

For the formation of a depression at the polar front, an essential condition is that the warm air should be moving to the eastward at a greater speed than the cold air. We may therefore have either of the two situations shown in Figs. 9.7*a* and 9.7*b*.

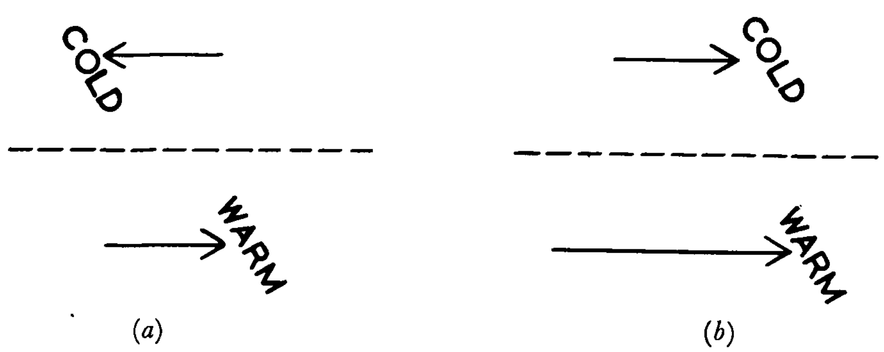

Fig. 9.7. Formation of a depression

In general, such a depression begins as a small wave-like disturbance on a frontal surface. As it develops a circulation it becomes a larger system and moves away to the E or NE in the northern hemisphere. (Figs. 9.8*a–d* represent successive stages in the development of a depression.) The warm air overrides the cold air at the warm front; the cold air undercuts the warm air at the cold front. Simultaneously a fall of pressure occurs over the centre; in other words, the depression deepens. Along with the process of development, there is a general motion of the system as a whole with the approximate direction and speed of the warm air. Thus we see that the polar front zone is the 'breeding ground' for temperate-zone depressions. Each of these depressions has its own warm and cold fronts which it retains for at least some of its journey across the ocean.

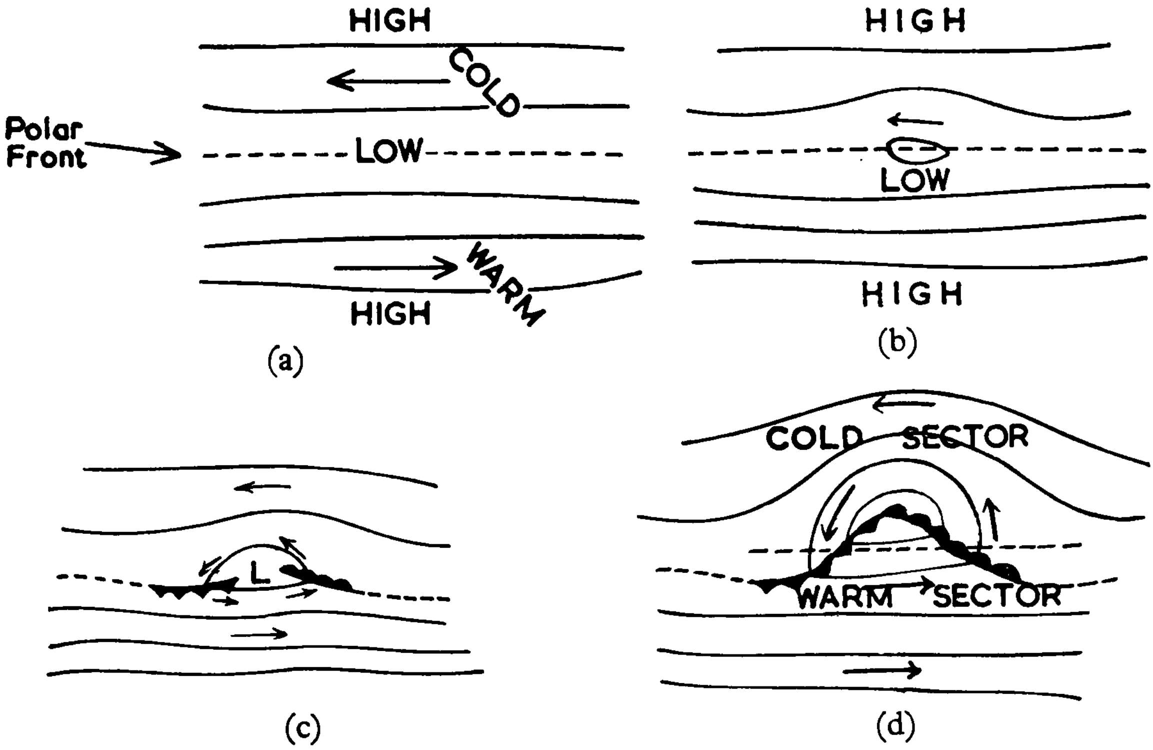

Fig. 9.8. Four stages in the development of a depression

Warm Fronts

When a warm air mass replaces a cold one, the line on which the frontal surface meets the ground is known as a WARM FRONT. The warm air overlies the cold air, which remains as a narrow wedge in contact with the ground. Earlier in this chapter it was explained that upward motion in the atmosphere is associated with horizontal convergence. At a warm front the warm air is flowing up the frontal surface over a wide area, and it is this extensive upsliding of air which is associated with the convergence of the warm air leading to the sequence of clouds and precipitation shown in Fig. 9.9.

Sequence of Clouds and Weather at a Warm Front

Fig. 9.9 gives the weather sequence at a typical warm front in the northern hemisphere. The vertical scale is much exaggerated, since in practice the angle of slope is only of the order of $\frac{1}{100}$ to $\frac{1}{200}$.

An observer in a position on the right of Fig. 9.9 will normally observe the following cloud sequence as the front approaches him: cirrus, cirrostratus, altostratus, nimbostratus. The rain normally commences to fall from the altostratus. After a steady thickening and lowering of the nimbostratus, patches

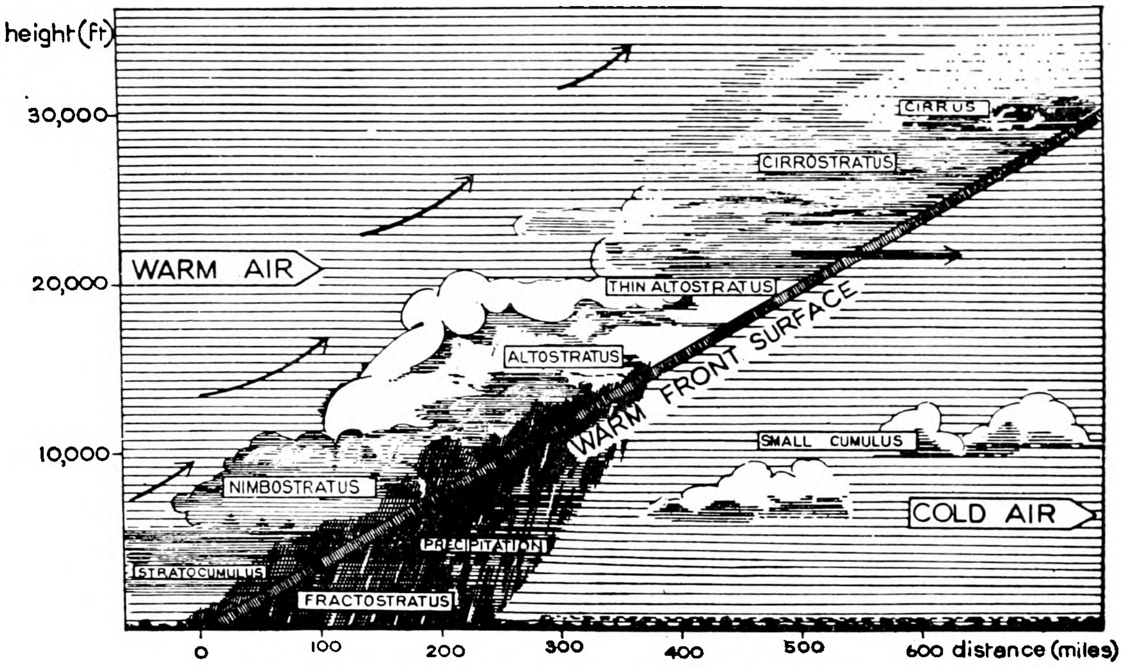

Fig. 9.9. Vertical section through a warm front

of stratus fractus or fractonimbus due to turbulent mixing in the cold air will appear, increasing as the front passes. Whether cumulonimbus will form depends on the stability of the warm air mass. The surface wind veers as the front passes in the northern hemisphere, usually becomes less gusty and decreases in force.

Table 9.1 gives a summary of the changes of different weather-elements at the passage of a warm front.

Table 9.1. Sequence of weather at a warm front

Element	In Advance	At the Passage	In the Rear
Pressure	Steady fall	Fall ceases	Little change or slow fall
Wind (northern hemisphere)	Increasing and sometimes backing a little	Veer and sometimes decrease	Steady direction
Temperature	Steady or slow rise	Rise, but not very sudden	Little change
Cloud	Ci, Cs, As, Ns in succession; scud below As and Ns	Low Ns and scud	St or Sc
Weather	Continuous rain or snow	Precipitation almost or completely stops	Mainly cloudy, otherwise drizzle, or intermittent slight rain
Visibility	Very good except in precipitation	Poor, often mist or fog	Usually poor; mist or fog may persist

Warm Sector

After the passage of the warm front comes the WARM SECTOR, the portion of the depression where warm air is in contact with the earth's surface, which is recognizable on the synoptic chart by nearly straight isobars. At sea it usually gives cloudy conditions, mainly low stratus or stratocumulus accompanied by occasional light drizzle. Visibilities are moderate or poor, and fog may occur.

Near the tip of the warm sector, where occlusion (*see* page 114) is taking place, thick cloud often with heavy rain occurs. In the regions of the warm sector remote from the centre of the depression and the tip of the warm sector, the weather becomes progressively better the further from the tip one goes; if one went far enough, anticyclonic conditions would eventually be reached.

Cold Fronts

A COLD FRONT is a line along which cold air replaces warm air. In this case, a blunt wedge or 'nose' of cold air pushes its way under a warm air mass which is thus forced to rise above the cold air. The cold air, being the denser of the two masses, remains in contact with the ground.

The slope of a cold front is much greater than that of a warm front, usually being about $\frac{1}{50}$. Consequently the upcurrents are more violent and cumulonimbus often appears. Here again there is a convergent wind-field in the lower layers which produces this vertical motion.

Sequence of Clouds and Weather at a Cold Front

Fig. 9.10 shows the typical sequence of clouds and weather at a cold front. The vertical scale is much exaggerated, the angle α being about $\frac{1}{50}$.

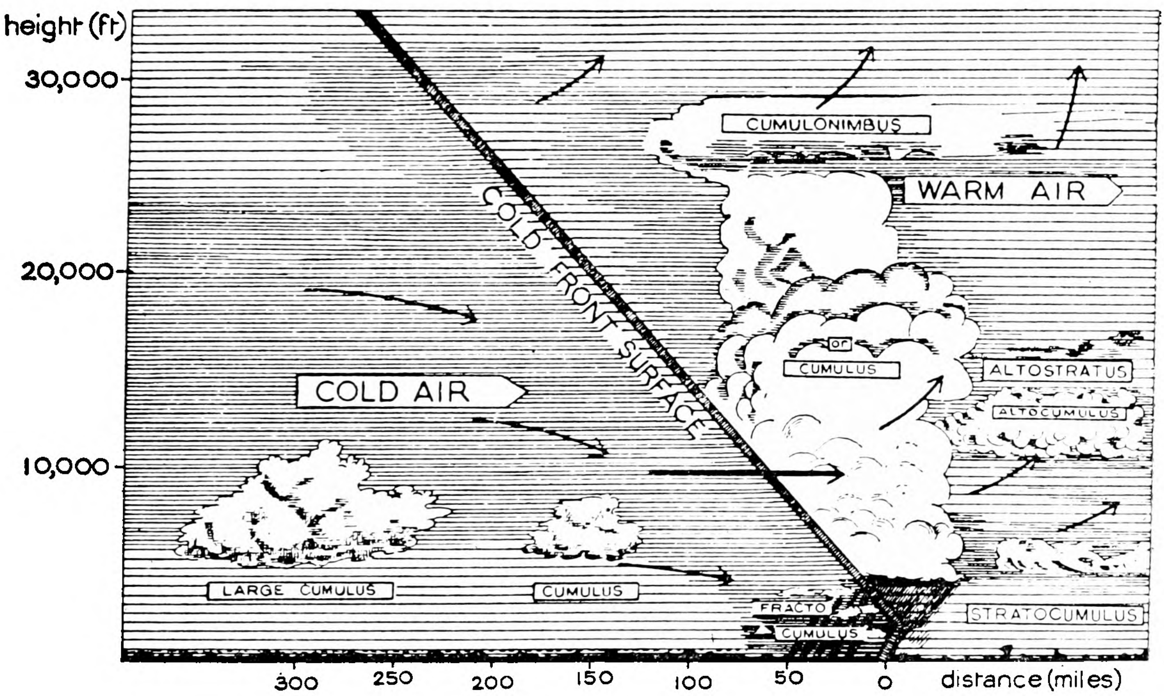

Fig. 9.10. Vertical section through a cold front

Provided gaps in the low cloud layers (stratus or stratocumulus) ahead of the front are large enough, an observer in a position on the right of Fig. 9.10 will observe cirrocumulus or patchy cirrus followed by altocumulus, slowly thickening into altostratus and then cumulonimbus. At the front, cumulus or cumulonimbus predominates and heavy rain for a relatively short period is typical; the surface wind veers quite quickly in the northern hemisphere and usually becomes squally and increases somewhat in force.

The diagram refers to a case where the upper wind is blowing decidedly across the front. In such a case there is a downward component of velocity over the upper part of the frontal surface. This prohibits any formation of upper cloud, and the cloud clearance after the frontal passage is very sharp. In cases where the upper wind is almost along the line of the front, there is no downward velocity along the upper part of the frontal surface; the frontal upper cloud structure therefore extends further to the rear, and the clearance after the frontal passage is much delayed.

Once the frontal weather has passed, the kind of weather that will be experienced will depend entirely on the character of the cold air mass. If this is unstable it will be characterized by cumulus or cumulonimbus cloud and occasional showers.

Table 9.2 gives a summary of the changes of different weather elements at the passage of a cold front.

Table 9.2. Sequence of weather at a cold front

Element	In Advance	At the Passage	In the Rear
Pressure	Fall	Sudden rise	Rise continues more slowly
Wind (northern hemisphere)	Increasing and backing a little, often becoming squally	Sudden veer and sometimes heavy squall	Backing a little after squall, then often strengthens and may steady or veer further in a later squall
Temperature	Steady, but fall in pre-frontal rain	Sudden fall	Little change or perhaps steady fall; variable in showers
Cloud	Ac or As, then heavy Cb	Cb with low scud	Lifting rapidly, followed by As or Ac; later further Cu or Cb
Weather	Usually some rain; perhaps thunder	Rain, often heavy, with perhaps thunder and hail	Heavy rain for short period but sometimes more persistent, then mainly fair with occasional showers
Visibility	Usually poor	Temporary deterioration followed by rapid improvement	Usually very good except in showers

Occlusions

As the depression progresses on its journey, the upsliding of the warm air at the warm front and the undercutting by the cold air at the cold front gradually diminish the extent of warm air at the surface, the latter being ultimately lifted from the ground and raised to greater altitudes. This shutting-off of the warm air from the ground is known as OCCLUSION. When the process has finished, the depression is said to be OCCLUDED. On the weather map this is shown by the cold front moving faster than the warm front and catching up with it, first near the centre where the fronts are close together and then at successively greater distances from the centre. Figs. 9.11*a–c* show successive stages in the occlusion of a depression. The line at the surface dividing the cold air which was previously ahead of the warm front from that which was previously behind the cold front is called an occlusion.

Most of the depressions which reach north-west Europe from the Atlantic are already occluded, and thus a large proportion of the fronts arriving over this

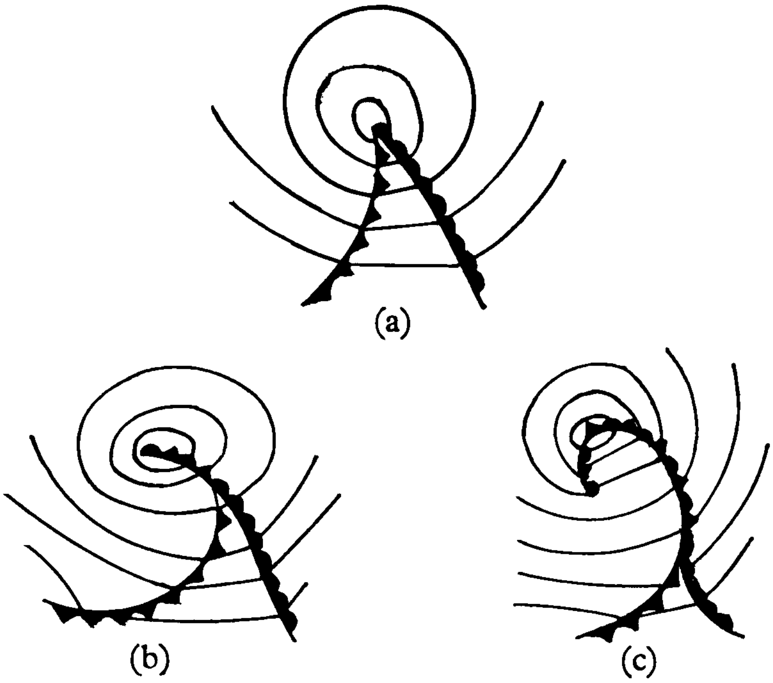

Fig. 9.11. Three stages in the development of an occlusion

area are occlusions. This was one reason why the acceptance of the Bjerknes idea on the development of temperate zone depressions was so long delayed, since the existence of warm sectors was not at first evident.

It is unlikely that the air masses on each side of the occlusion will have identical properties in view of their different history, so that the air following the occlusion may be either warmer or colder than the air preceding it. In the first event, the occlusion is said to be of the warm-front type, or simply a WARM OCCLUSION; in the second case, the occlusion is of the cold-front type (COLD OCCLUSION).

Figs. 9.12*a*, *b* show examples of warm and cold occlusions respectively in section. Fig. 9.12*a* shows what happens when the cold air in the rear of the occlusion is less cold, and therefore less dense, than the cold air ahead of it. Both the warm air and the less cold air override the preceding cold wedge. The discontinuity between the warm air and the less-cold air is still to be found aloft, and the line in which this surface meets the original warm-front surface is referred to as an UPPER COLD FRONT. The upper cold front is marked by cumulonimbus clouds and rain of a showery type, which are superimposed on the normal warm-front distribution of weather. The wind veers at an occlusion in the northern hemisphere; usually the veer is more marked at a cold occlusion than at a warm occlusion.

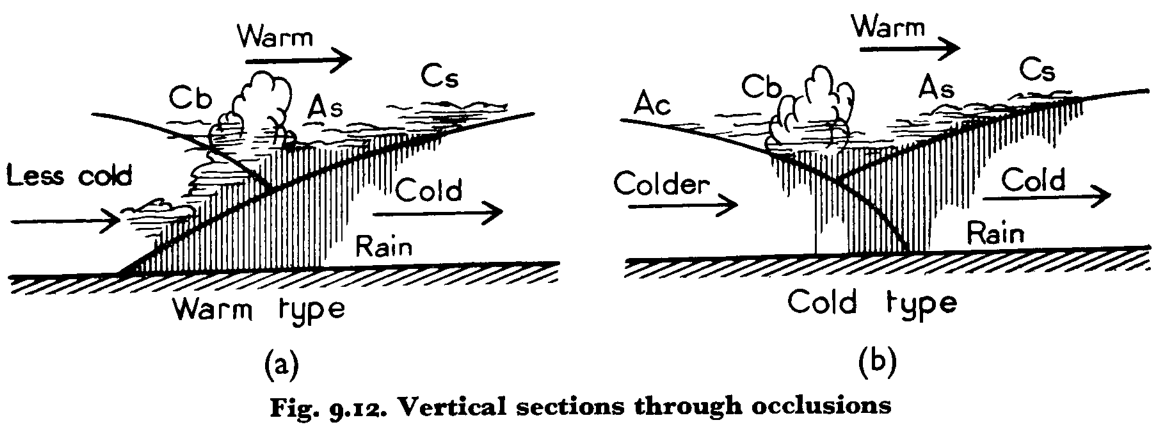

Fig. 9.12. Vertical sections through occlusions

Referring to Fig. 9.12*b*, the cold air in the rear of the occlusion, being colder and denser than the cold air ahead of it, necessarily remains in contact with the ground and constitutes the under-cutting wedge, obliging the less-cold air ahead of it to rise. The discontinuities between the two types of cold air and the warm air still exist aloft but are no longer apparent at the ground.

Ahead of the occlusion the weather characteristics are those of a warm front, since the warm air continues to rise over the cold wedge of air. At the occlusion the wind veers in the northern hemisphere, as it does at the passage of all fronts, but there is no rapid clearance; there is only a belt of more intense rainfall with heavy cumulonimbus cloud formed by the upthrusting of both warm and cold air masses by the colder following wedge. Squally conditions with an increase of wind in the colder air also occur frequently.

Occlusions which penetrate into the cold continents from the oceans in winter are generally of the warm type, because maritime-polar air is warmer than continental-polar air. In the summer, cold occlusions bring cold maritime air into the relatively warm continents.

Filling Up and Dissolution of Depressions

Once a depression is fully occluded it will usually fill up and will no longer be recognizable on a synoptic chart within a few days. When there is little horizontal contrast of temperature between the air in different regions of a depression it usually becomes almost stationary, and this is also true when air which is relatively cold for its level is located symmetrically over the centre. Sometimes a depression which has almost filled up disappears by being absorbed into the circulation of a new system; on other occasions a completely occluded depression is observed to deepen. When this happens, it is generally found to be associated with the arrival of a new supply of cold air or, less commonly, of warm air, which is drawn into the circulation of the system as a result of changes in the pressure distribution in the surrounding regions. In a well-occluded depression, that is a depression in which the occlusion has almost disappeared and can no longer be recognized on the chart by any belt of medium cloud or precipitation, the weather is marked by showers and bright intervals with generally good visibility in the polar air. In the last stages the showers become rather scattered and infrequent.

General Distribution of Weather in a Warm-Sector Depression

Different aspects of the depression have been so far discussed separately. Fig. 9.13 shows the distribution of cloud and weather in a typical depression in plan, and Figs. 9.14*a*, *b* show sections of it taken north and south of the centre respectively.

Fronts in Relation to Isobars and Wind

Fig. 9.15*a* shows a frontal surface and two vertical columns of unit cross-section; column **W** entirely in the warm air, column **C** partly in the warm and partly in the cold air, while the pressures at the tops of the columns are equal. Since the cold air is denser than the warm, it follows that column **C** exerts a greater pressure at the surface than does column **W**.

Now consider Fig. 9.15*b*, which represents the atmospheric pressure at the ground. Proceeding along an isobar (say 1004 mb) from **A** to **B**, on reaching the front we enter the cold air where the pressure increases on account of the extra

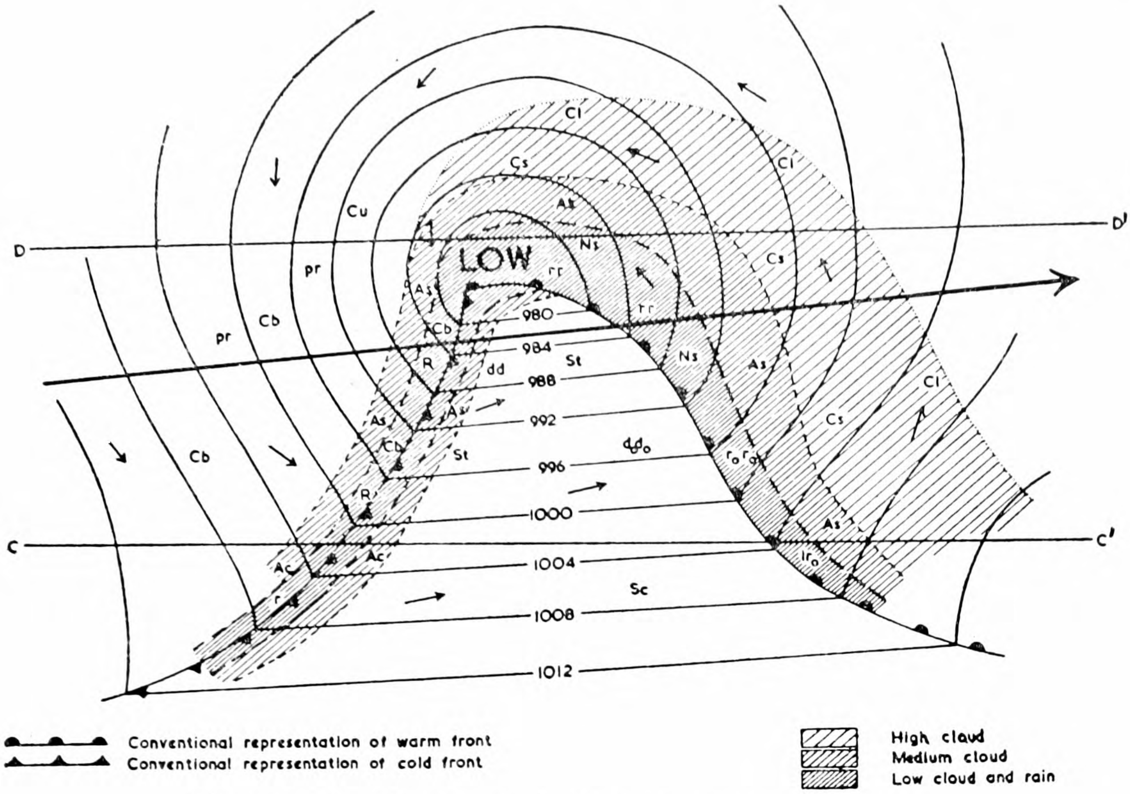

Fig. 9.13. Weather in a typical warm-sector depression

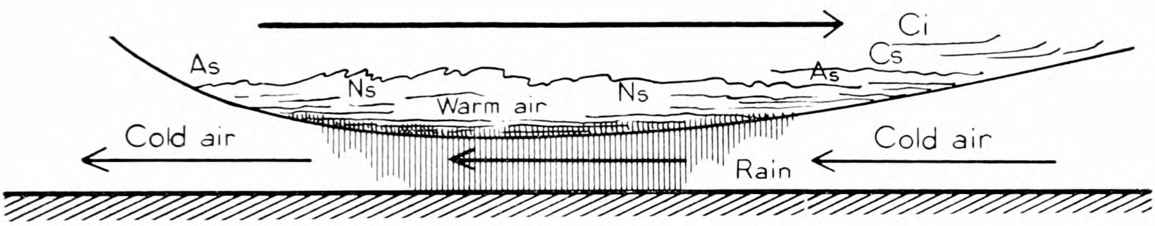

Fig. 9.14a. Section of typical depression along DD′ north of centre

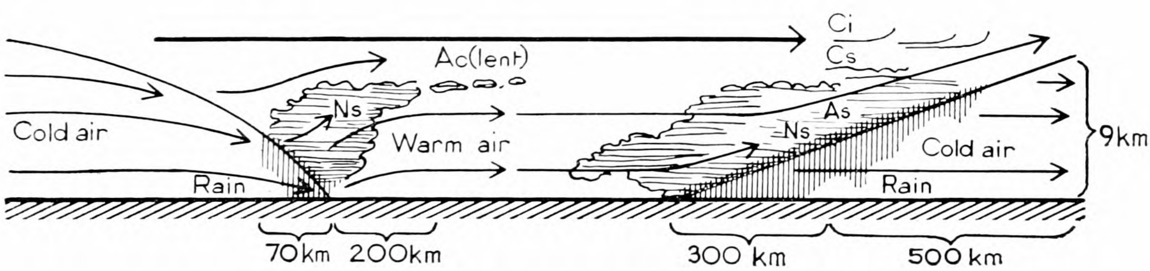

Fig. 9.14b. Section of typical depression along CC′ south of centre

weight of the colder air; if we continue in the same direction far enough we meet the 1008 mb isobar. This shows that the isobars at a front are refracted or sharply bent in such a manner that the kink in the isobar points from low to high pressure. This characteristic applies to all types of front.

The pressure will fall as the front approaches a place and rise, or fall less rapidly, after its passage. As the front passes the barograph shows a sharp downward kink, its magnitude and form depending on the sharpness and speed of the front.

From the relation of the front to the isobars we can find its relation to wind, emembering that the wind near the earth's surface blows mainly along the

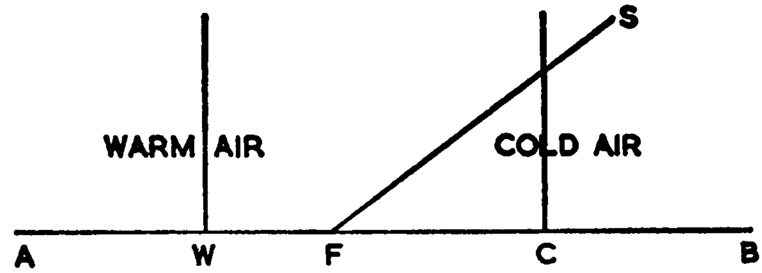

Fig. 9.15a. Vertical section along AB in Fig. 9.15b

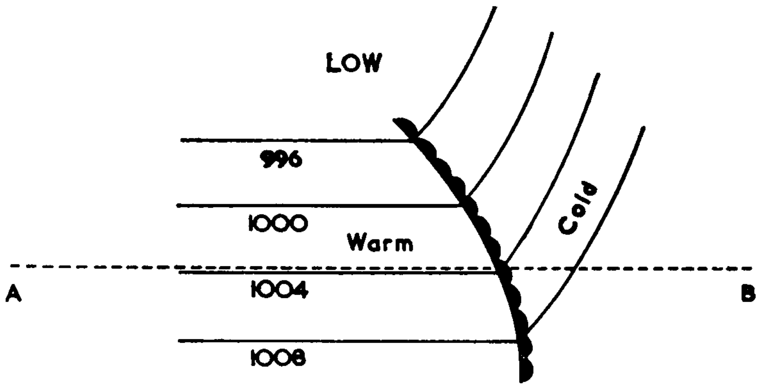

Fig. 9.15b. Isobars at a warm front

isobars with a slight deviation towards the side of lower pressure. The kink in the isobars causes the sudden veer of the wind which occurs at the passage of a front. The following rule holds for all types of front in the northern hemisphere:

"Face the wind in advance of the front and the wind will shift to the right as the front passes."

Families of Depressions

A disturbance forms at the polar front and in its early stages travels along it. It is useful to regard the fronts of the depression as being just a distortion of the main polar front (Fig. 9.8).

Coincident with the formation of one depression, conditions are usually favourable for the formation of others, which travel in a series or family along the main frontal zone, the cold front of each depression being swept back to form the warm front of the succeeding one. As each depression deepens, its cold front is swept further and further southwards so that succeeding members of the family tend to form further south than the preceding one. Eventually the cold air behind one of the depressions sweeps through to the trade winds and the series is broken, the next depression forming much further north on a regenerated polar front.

The number of depressions in a family varies, but averages about four. The depressions are normally separated by ridges of high pressure which give brief, fair intervals between the rainy periods.

Secondary Depressions

By 'secondary' is meant a depression embedded in the circulation of a larger or more vigorous depression, known as the primary. (*See* Fig. 2.2.) In general, the secondary moves around the primary in a cyclonic direction. When the primary is weak and the secondary strong, as often occurs during the 'filling-up' stage of the primary depression after it has become well occluded, the

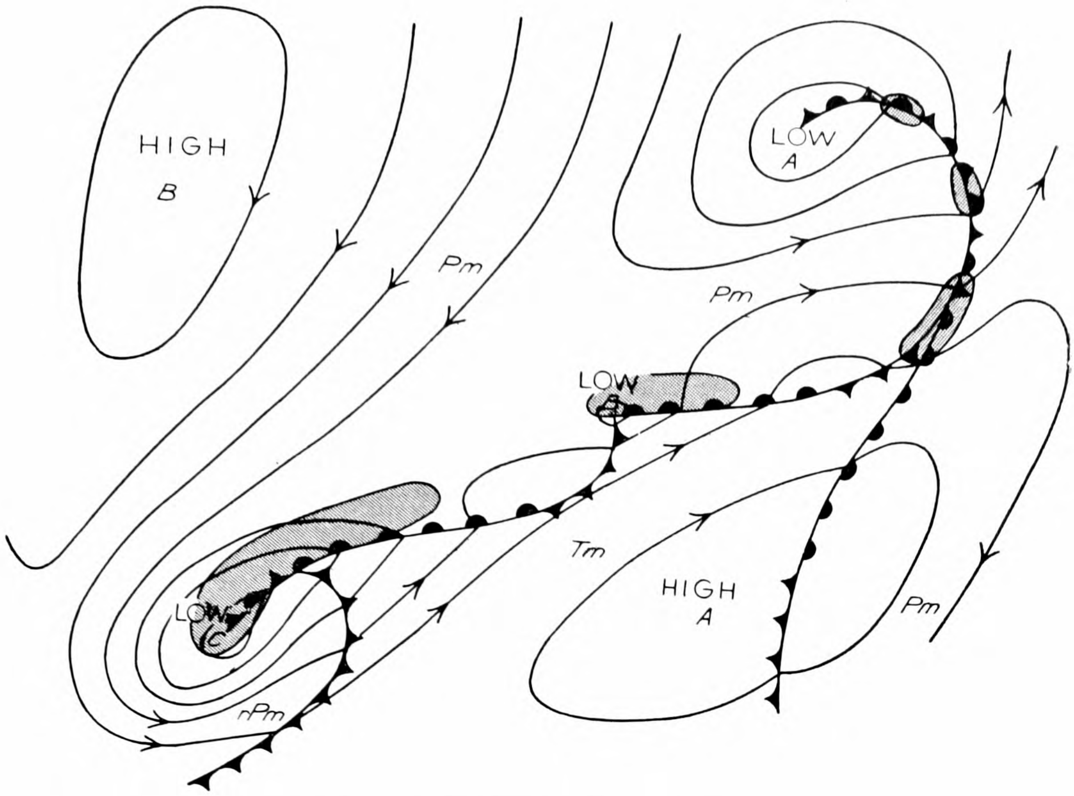

Fig. 9.16. A family of depressions

Low *A*: old occluded low.
Low *B*: cold-front wave.
Low *C*: slow-moving low at low latitudes.
Also, High *A* is part of the subtropical high-pressure belt and High *B* is a slow-moving blocking high.
Shaded areas represent rain areas.

primary may be absorbed in the circulation of the secondary as described earlier, or the two depressions tend to rotate around each other in a cyclonic sense. When the depressions have formed on the polar front, the secondary may be simply the next member of the family which has moved quickly and been drawn into the more vigorous circulation of the preceding member.

Non-frontal Depressions

Most depressions of temperate latitudes form on the polar front, and for this reason emphasis has been placed in the foregoing paragraphs on the formation of these 'frontal' depressions. Other types of depression exist, however, which are not connected with frontal zones. They are as follows:

(*a*) Thermal depressions.
(*b*) Depressions due to vertical instability.
(*c*) Depressions due to topography ('Lee depressions').

Thermal Depressions

The formation of thermal depressions is due to unequal heating of adjacent surface areas, and land and sea distribution plays a big part in determining their location. In winter the cooling of the continents induces higher pressure over the land than over the sea. In summer pressure tends to be lower over the land than over the sea. This effect may be seen by comparing Figs. 7.1 and 7.2 showing world pressure distribution for January and July. The low pressure over Asia in July, known as the South Asiatic Monsoon Low, is pronounced enough to control the atmospheric circulation over a vast area.

Examples on a smaller scale are shown by the occurrence of low pressure over inland seas in winter, e.g. Mediterranean, Black Sea, Caspian Sea.

Depressions due to Vertical Instability

Vertical instability probably plays a part in the deepening of all types of depressions. It has been mentioned that low pressure tends to occur over inland seas in winter due to unequal cooling of land and sea. It often happens, especially in the Mediterranean, that these depressions deepen much more than would be expected from the effect of differential heating between land and sea. One reason for this deepening is the vertical instability that arises when continental-polar air flows over a warm water surface, which results in the formation of rather prolonged and widespread showers or precipitation.

An instability depression sometimes develops entirely within a mass of polar air flowing towards lower latitudes over a progressively warmer sea surface which provides heat and moisture. The steep horizontal temperature gradient produced along the air flow favours the formation of such a depression. At first the system may only show as a bulge in the isobars; later it may develop into an extensive system with closed isobars.

Vertical instability is important in the formation of tropical cyclones and in much smaller rotating systems such as tornadoes, dust devils and waterspouts.

Depressions due to Topography ('Lee depressions')

In a few localities where the winds sometimes blow across a mountain range which is sufficiently high and continuous to act as a barrier, the resulting distortion of the wind flow leads to the formation of a depression in the lee of the mountain range. (*See* Fig. 9.17.) A depression formed in this way is known as a LEE DEPRESSION. Sometimes such a depression deepens and moves away, later becoming indistinguishable from a frontal depression.

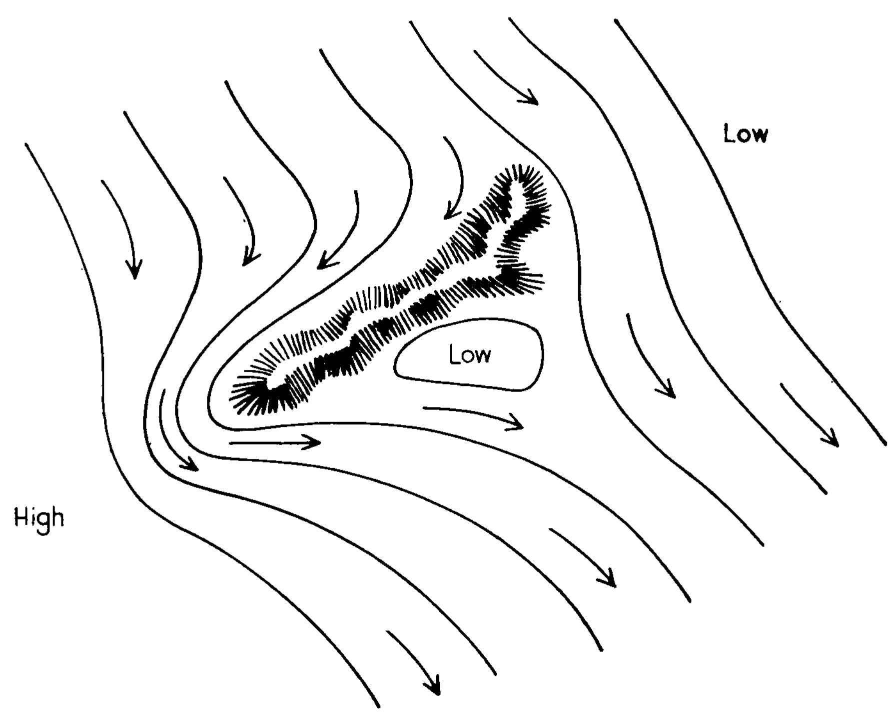

Fig. 9.17. A lee depression

Depressions are formed in this way in the Gulf of Lions, when the N to NW wind known as the Mistral blows down the Rhône Valley, after passing through the gap between the French Alps and the Cevennes. This wind usually occurs in the rear of a cold front moving SE across the region and often reaches gale force. The resulting lee depression often moves NE from the Gulf of Lions to the Lombardy Plain and into Central Europe. The formation of shallow, mostly slow-moving depressions in Central China, which are of great importance there on account of the rainfall they produce, is attributed to distortion in the flow of NW winds when crossing mountain ranges in their path. Stationary lee depressions caused in the same way are often found to the west of Madagascar and in some other parts of the world.

Depressions—General Remarks

The birth, development and dissolution of a depression are only stages in its life history; its development is the product of environment, which it can modify. Thus the circulation of a depression, by drawing different air masses together, can contribute to its own deepening.

The Bjerknes theory of frontal depressions provides a model or pattern whereby the development of any depression can be assessed. Experience has led to the formulation of rules, some of which have some support in theory, e.g. that a depression moves in the direction of the isobars in the warm sector.

Newly formed depressions tend to follow the circulation in which they are formed. Thus a secondary depression will tend to move cyclonically around its primary. This is the 'steering' principle whereby a pressure system is 'steered' along the circulation of a larger system. It is also found that the circulation pattern aloft has steering properties, and this fact is used in forecasting when upper air charts are available.

As a depression develops, or deepens, it grows in size, as shown by the increasing number of closed isobars on the synoptic chart. While some portion of the warm sector remains at ground level, a depression will continue to move with a speed which is about four-fifths of the geostrophic wind in the warm sector. The most usual direction of travel is nearly the same as the geostrophic wind in the warm sector, but inclined at a small angle to it away from the low pressure. After occlusion it will move more slowly or become stationary, depending upon the intensity of the surrounding pressure field at the surface, and the magnitude of the horizontal thermal gradient at different levels in the troposphere. A really deep depression may include a considerable part of an oceanic area in its circulation. Depressions with central pressures under 960 mb may be regarded as really deep. Pressures of less than 930 mb at the centre of depressions have been recorded on very few occasions. The deepest depressions have complex life histories and cannot be explained solely by a simple occlusion process. They can draw air masses from different sources into their circulations, thus providing a mechanism for deepening by stages.

The process whereby a new depression forms, or an existing one deepens, is known as CYCLOGENESIS. Since cyclogenesis is associated with strong horizontal temperature-gradients, the principal areas of its occurrence are found in temperate latitudes off the eastern coasts of the continents, particularly in winter.

Thermal Winds

The reader who is not interested in the meteorology of the upper air can omit these paragraphs which are included in order to give a fuller understanding of the relationship between winds and pressure distribution at all levels.

In a region where the isobars are straight and parallel and the pressure gradient is uniform, the geostrophic wind, being directly proportional to the magnitude of the pressure gradient, will have the same magnitude over the whole region. In addition, if the lapse rate is the same everywhere over this pressure field, it follows that the horizontal differences of temperature will be the same at all levels over the area as they are at the surface. In practice this state of affairs is never found to exist in the atmosphere; on the contrary, at higher levels in the troposphere the air is often found to be colder over a depression than it is at the same levels outside the depression, and similarly the air at the centre of an anticyclone is often warmer than at the same levels over surrounding areas. We shall now consider in more detail what effect is produced upon the pressure and wind distribution aloft by the presence of colder air at higher levels in the middle of a depression, and correspondingly by the presence of warmer air at higher levels in an anticyclone.

The effect can be illustrated very simply. Suppose we consider three columns of air **A**, **B** and **C** (as in Fig. 9.18), where the bases of all three columns have the same surface pressures and surface temperatures. The lapse rates in columns **A** and **C** are also the same, so that the temperatures and pressures in these two columns have the same value at corresponding levels above the surface. Now suppose that the lapse rate in column **B** is steeper than in columns **A** and **C**; that is to say, at higher levels the air in this column is colder that the air at corresponding levels in columns **A** and **C**. It is readily inferred that the pressure in the colder column **B** will decrease more rapidly with height than the pressure

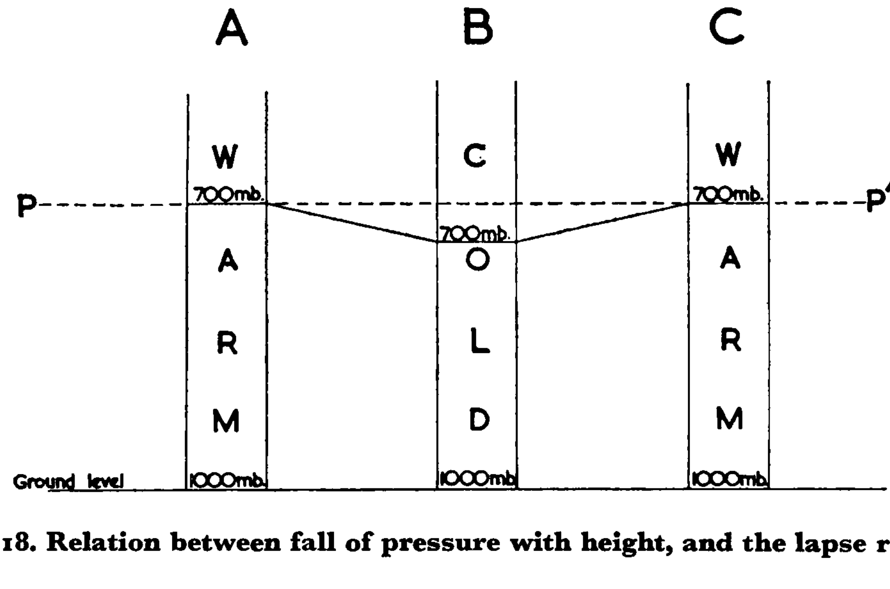

Fig. 9.18. Relation between fall of pressure with height, and the lapse rate

in the other two columns. As a result we have the situation shown in Fig. 9.18. Here, for the sake of simplicity, the pressure at the ground level is everywhere shown as 1000 mb. At the level **PP′**—since the lapse rates in columns **A** and **C** are equal—the pressures in columns **A** and **C** will have the same value (say 700 mb), but will have some value *below* 700 mb in column **B** at this level. From these considerations it can be understood why a synoptic chart drawn for a higher level (for instance, 10,000 ft) will show an area of low pressure where the air is relatively cold. Another and more convenient method which makes use of the same result, is to draw synoptic charts which show the level of standard pressure surfaces (usually 700, 500, 300 and 200 mb levels), by means of isopleths joining together points along which the given pressure surface is found at the same height above the ground. These isopleths are equivalent to contour lines of the given pressure surface, and are normally drawn for height intervals of 200 ft or 60 m as routine practice in many meteorological services.

Buys Ballot's law relating wind and pressure gradient is independent of pressure, hence it follows that a cyclonic circulation is set up at higher levels around a cold column of air, and an anticyclonic circulation at higher levels around a warm column of air in the troposphere, and that these circulations are independent of the surface pressure gradients. The component of wind resulting from this cause alone is described as the THERMAL WIND. The upper wind at any level may conveniently be thought of as consisting of two vector components, namely, the 'geostrophic' and 'thermal' wind components; hence the variation of wind direction and speed with height is dependent upon the location of cold or warm air at higher levels (nowadays called COLD POOLS and WARM POOLS of air) in relation to the direction and horizontal gradient of the surface isobars. When the pressure and temperature gradients are parallel and the colder air above coincides with the low pressure, the wind increases with height; similarly when the pressure and temperature gradients are parallel but the warmer air lies over the low pressure the wind decreases with height; in both these cases the wind direction does not alter with height so long as the pressure and temperature gradients remain parallel. The case when the wind increases or decreases with height without change of direction is the simplest case of wind 'shear'.* As a rule the pressure and temperature gradients are not parallel, and this results in a change of wind direction in addition to the change of speed with height.

* Wind shear is a change of wind over a short distance, either vertically or horizontally.

CHAPTER 10

ANTICYCLONES AND OTHER PRESSURE SYSTEMS

Anticyclone

An anticyclone is defined as a region where a system of isobars encloses a region of high barometric pressure. Buys Ballot's law says that if you face the wind the lower pressure is on the right in the northern and on the left in the southern hemisphere—consequently the wind circulation in an anticyclone must be clockwise in the northern and anticlockwise in the southern hemisphere.

In the central areas of most anticyclones light winds and fair weather prevail. This is true for the subtropical high pressures and for anticyclones in temperate latitudes in summer. Although cloudy skies are somewhat frequent in anticyclones over the ocean in temperate latitudes, precipitation even as drizzle is rather uncommon near the middle of an anticyclone because active fronts do not penetrate such regions. In winter over the continents in temperate latitudes, and particularly near industrial areas, the weather in an anticyclone can seldom be described as 'fair'. The reason is that in these circumstances fog and 'smog' are common and often persist for some time in the lowest layers of the atmosphere, with serious effects upon the movement of shipping in estuaries and coastal waters.

We have already seen that, at the earth's surface, the air does not flow exactly along the isobars but is deviated slightly from high to low pressure. If this be applied to a closed system of isobars with high pressure on the inside, it is apparent that air is continually being lost from the area; in other words, the anticyclone is an area of horizontal divergence in its lowest levels. As explained earlier, horizontal divergence at the earth's surface implies a downward motion of the air, which is accompanied by subsidence and by adiabatic heating. Hence the relative humidity of air in the lower levels of anticyclones is lowered, clouds tend to be evaporated, and the frequent occurrence of fair weather in anticyclones is thus explained. Also subsidence often leads to the formation of an inversion of temperature at a low level in an anticyclone which has been in existence for several days.

Anticyclones are usually classified as 'cold' or 'warm'.

Cold Anticyclones

A cold anticyclone is one in which the air at the surface, and in the lower layers of the troposphere, is colder than the air in adjacent regions. The air in the anticyclone is thus denser than the surrounding air, level for level. The high pressure of a cold anticyclone is therefore due, primarily, to the density of the lower layers of the troposphere being greater than the density of the same layers in the area surrounding the anticyclone.

Examples are the 'semi-permanent' anticyclones over the continents in winter. These anticyclones most often occur over Siberia and less frequently over North America. They are not strictly permanent because these areas are occasionally invaded by travelling depressions; but after each period of cyclonic activity, high pressure tends to be re-established, and may then exist for weeks with little change.

These seasonal anticyclones are predominant enough to show on the winter chart of mean pressure (Fig. 7.1), and they control the atmospheric circulation over wide areas. They are the 'source-regions' of continental-polar air. Countries which normally enjoy mild maritime conditions, occasionally experience the rigours of a continental winter during an invasion of continental-polar or arctic air. In the British Isles, this sometimes occurs in winter months when a separate anticyclone develops over northern Europe, resulting in persistent easterly winds across the North Sea and the British Isles.

On the east coasts of North America and Asia the outflow of continental-polar air maintains a sharp temperature contrast off the coast, where a warm ocean current runs. These temperature contrasts favour the formation and development of depressions, which then travel eastwards.

Although cold anticyclones are of limited vertical extent (i.e. the height to which an anticyclonic circulation of air extends does not exceed 10,000 ft), they play an important part in the low-level atmospheric circulation in winter.

In addition to these seasonal cold anticyclones, other more transitory ones exist. The frontal depressions of temperate latitudes travel eastwards in families, the depressions being separated from each other by ridges of high pressure travelling with them and bringing the weather clearances that occur after the passage of each depression. After the last member of the family has formed, the cold polar air in its rear begins to sweep towards the equator and an anticyclone builds up in the cold air. Its history is then a matter of circumstance. If formed over land, in winter, the transitory anticyclone will be maintained and perhaps intensified, and eventually become an extension of the main semi-permanent anticyclone. If formed over the sea, or over the land in summer, the transitory anticyclone either collapses rapidly or becomes transformed into a warm anticyclone, the cold air being heated adiabatically by subsidence.

Warm Anticyclones

In a warm anticyclone, the air throughout the greater part of the troposphere is warmer, level for level, than its environment. Near the surface the air in a warm anticyclone therefore differs little in density from that of surrounding air. Surface pressure is dependent on the total mass of air above the surface where it is measured, hence the excess surface pressure in a warm anticyclone must result from air of greater density than the surrounding air at higher levels. This is brought about by convergence of air at these levels accompanied by subsidence at lower levels. In a warm anticyclone, pressure at higher levels is higher than in the surrounding air, level for level, since pressure falls more slowly with height through a column of warm air than through a column of cold air.

Examples of warm anticyclones are the oceanic subtropical belts of high pressure. (*See* Figs. 7.1 and 7.2.) They are an essential feature of the general circulation, and cover areas where the air is subsiding. Consequently the weather there is generally fine, with little or no cloud and good visibility. The subtropical anticyclones are the source-regions of maritime-tropical air masses, providing the warm air that feeds travelling depressions of temperate latitudes. The charts of mean pressure show the centres of these anticyclones as being located in lat. 30° to 35°N and S; in fact, in addition to showing a regular variation of mean position between seasons, their positions vary between corresponding seasons in different years. The causes of these movements are not yet known, variations in the strength of the equatorial ocean currents and of the amount of ice in the North Atlantic are among many other factors which have been suggested to account for them, but convincing proof of any such relation remains to be found. However, there is no doubt that these variations have an important effect on the seasonal weather in many localities. For example, in summer the Azores anticyclone lies farther north than in winter, so that fewer depressions affect the British Isles in summer than in winter, and therefore summer is less windy and rainy on the whole than winter in these islands. Most spells of dry warm weather in the British Isles result from occasional NE extensions or offshoots of the Azores anticyclone taking up a position over the British Isles; and from time to time such spells may last with little interruption for weeks on end.

Upper air soundings made above anticyclones show that the air is relatively dry, which is to be expected in view of the subsidence taking place. They may also show an inversion of temperature, known as a 'subsidence inversion' because it is formed as a result of subsidence, which is most effective in a layer a short distance above the ground. As a result the highest temperature in an anticyclone is often at a level of about 1,000 to 2,000 ft above the surface; this is particularly true over the sea. To some extent this inversion aids fog formation. At sea this happens because the moisture evaporated from the sea is all kept in the lowest layers, particularly if the airstream is moving towards higher latitudes, that is towards progressively colder sea surfaces. Over land other factors, such as radiation to clear skies and high moisture content, are assisted by the absence of wind associated with the inversion to produce a fall of temperature enough to cause fog by condensation.

Clouds forming in the lower layers of an anticyclone, as, for example, when moist air from the sea flows over land that is being heated by the sun, tend to spread out at the temperature inversion, giving a layer of stratocumulus. This is more typical of the boundaries of an anticyclone than the centre, where the subsidence frequently prevents cloud formation.

Straight Isobars

This phrase is used to describe isobars which are nearly straight and parallel over an area for a few hundred miles. Although straight isobars are not usually classified as an independent pressure type, it is convenient to consider them before other pressure types. There is no well-defined weather associated with straight isobars. The weather depends on the source of the air mass and its subsequent history. With straight isobars it is relatively easy to follow the track of the air and hence deduce the properties of the air mass, and the resulting weather.

An example is one in which isobars run north to south across the British Isles from the Norwegian Sea, pressure being high to the west and low to the east. The air must move south across a progressively warming sea surface, and therefore becomes unstable and the weather showery. The clouds are cumulus and cumulonimbus, sometimes massed together, but usually with appreciable breaks.

With straight isobars the type of weather will also depend on the steepness of the pressure gradient; thus the isobars may be straight in the outer parts of an anticyclone or in the warm sector of a depression, and if the depression is deep gales will occur.

Trough of Low Pressure

The isobars in a trough of low pressure resemble the contours round a river valley or a submarine valley or canyon. Pressure is lowest within the 'V'- or 'U'-shaped isobar of lowest value. Frequently a front lies in the trough, and the isobars may show a marked kink or bend at the front. A number of troughs do not contain fronts; these are known as non-frontal troughs. They are usually well rounded and do not show any sharp bend in the isobars such as occurs at a front. Figs. 10.1*a*, *b* illustrate both types of trough. A non-frontal trough typically occurs in the cold air mass behind an occluded depression, and the resulting increase of pressure gradient often leads to a renewal of strong winds or gales after the depression centre has passed.

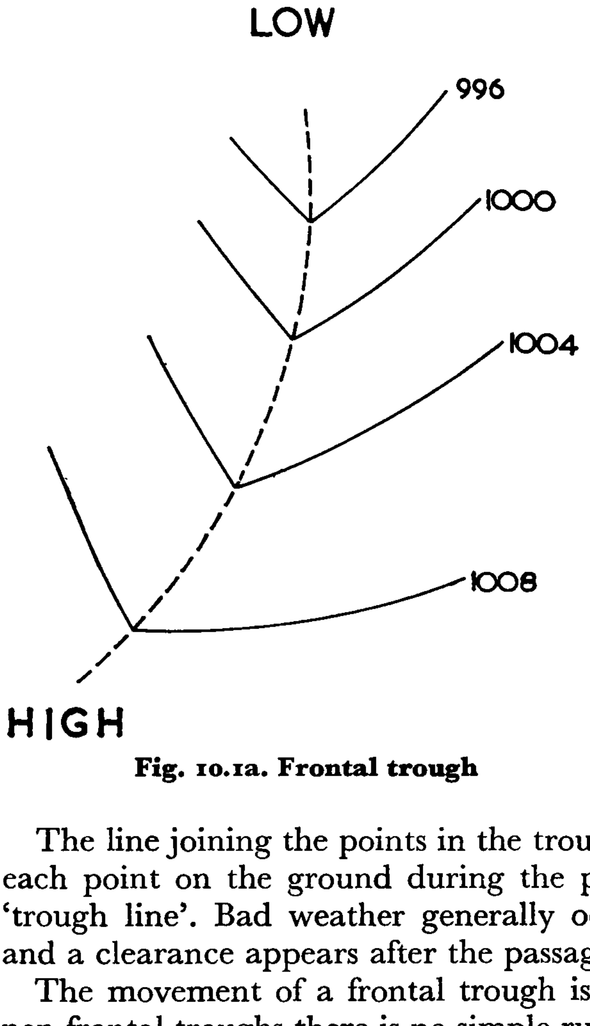

Fig. 10.1a. Frontal trough

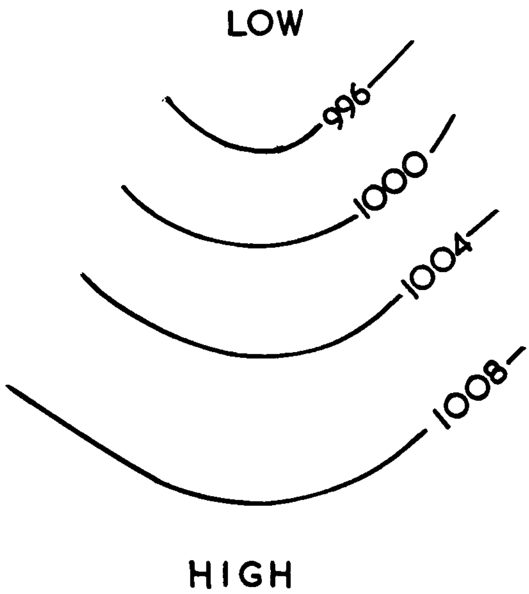

Fig. 10.1b. Non-frontal trough

The line joining the points in the trough where the lowest pressures occur at each point on the ground during the passage of the trough is known as the 'trough line'. Bad weather generally occurs on the advance side of troughs and a clearance appears after the passage of the trough line.

The movement of a frontal trough is the same as that of its front, but for non-frontal troughs there is no simple rule. When associated with a depression, they are swept along in its circulation and rotate around its centre.

Some troughs tend to remain stationary for long periods in temperate latitudes, particularly in summer when the upper air circulation pattern is weak, i.e. when winds are light at most levels. In these circumstances weather is mostly cloudy with local showers or thunderstorms, which may be heavy near an active front. In tropical latitudes, low-pressure troughs are regions favourable for the development of tropical storms in certain ocean areas.

Ridge of High Pressure

A ridge of high pressure (Figs. 2.1 and 2.3) is a system of curved isobars in which pressure is higher on the inside than the outside. It therefore has anticyclonic properties. The isobars in a ridge are not always closed. Ridges are often located in the outer regions of anticyclones; and weather conditions in a ridge are often very like those in an anticyclone.

Fast-moving ridges occur between individual members of a depression family, moving along the polar front in temperate latitudes. They bring intervals of fair weather between the periods of rain or showers associated with each passing depression and its fronts. With a moving ridge a progressive deterioration of weather is usual after the passage of the axis of the ridge. When depressions follow one another a short distance apart, as often occurs in the Southern Ocean and in winter in the North Atlantic, the duration of the fair interval may be a very few hours.

Ridges formed as offshoots from the sub-tropical anticyclones, or from the semi-permanent winter anticyclones of the continents, usually move fairly slowly.

Fine weather is associated with a ridge of high pressure, although the qualifying circumstances mentioned for anticyclones also apply with ridges.

Col

A col is the region between two ridges of high pressure and two troughs of low pressure situated alternately. (*See* Fig. 2.4.)

No definite weather can be associated with a col; indeed it is usually a region where sharp changes occur. The reason is not difficult to see. Suppose we cross the col with two diagonal lines **AB** and **CD** (Fig. 10.2), **AB** is a line along which

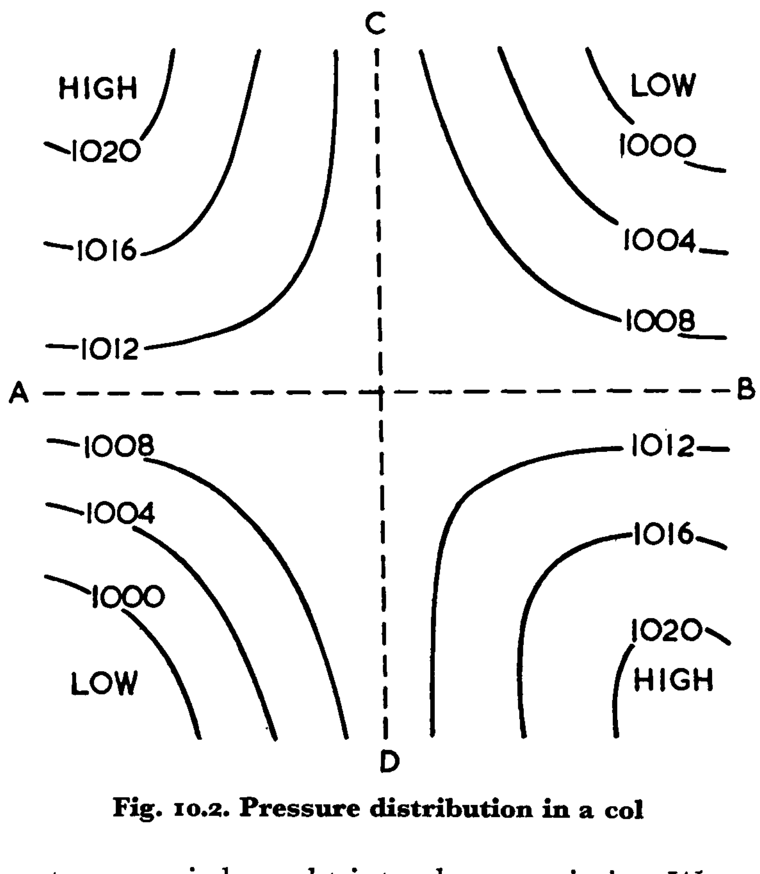

Fig. 10.2. Pressure distribution in a col

air from different sources is brought into close proximity. We might therefore expect a front to occur somewhere near this line, and this often occurs in practice. Along **CD**, on the other hand, the adjacent air flow is not front-forming. Thus a front can often be found in a col and it must lie along or close to the line **AB**. A front is not always found in a col, however, because its formation requires favourable conditions over a period of time.

The centre of a col is associated with very light and variable winds, and thus it is an area in which fog may occur, particularly in autumn and winter over land. When a col lies over a land area in summer, thunderstorms commonly occur when the necessary moisture and instability are present at higher levels.

CHAPTER 11

TROPICAL REVOLVING STORMS

General Remarks

A tropical revolving storm may be defined as a small area of intense low pressure around which winds of hurricane force blow, inclined slightly towards the centre, in an anticlockwise direction in the northern hemisphere and in a clockwise direction in the southern hemisphere. These storms as a whole (or the centre of the storms) have a forward or progressive motion and often produce mountainous seas and weather conditions of a far more devastating character than the worst storms of temperate latitudes.

Description of Tropical Revolving Storms

SHAPE. Pressure gradients are generally much weaker in low than in high latitudes. When a tropical revolving storm is formed, however, the gradient becomes steep, often extremely so. The deflecting force due to the earth's rotation is very small in the low latitudes in which tropical storms form and the steep gradient, which 'pushes' or 'pulls' the air inwards towards the low-pressure centre, is balanced by the centrifugal force of air revolving in a small circle, which pulls it outwards from the centre. As a result the isobars of tropical storms resemble true circles.* This symmetry, inherent, oddly enough, in one of the most violent manifestations of nature, simplifies the 'laws of storms' as described in *Admiralty Pilots* and other navigational works. These laws aim at getting the ship out of the storm field as quickly as possible, and while she remains in it, avoiding the dangerous area near the centre where the wind and violent squalls are strongest and the seas highest and most confused. (*See* page 145.)

While the shape of the isobars may be circular, they are not necessarily evenly spaced and there is frequently a crowding together on the poleward side in the direction of the anticyclone around which the storm is travelling. As a result the strongest winds may be expected in this semicircle.

SIZE. From the steepness of the pressure gradient, it follows that a tropical storm will generally cover a smaller area than a temperate depression. Whereas the average diameter of a temperate depression is 1,000 to 1,200 miles, that of a tropical storm is often less than 100 miles. The largest tropical storms rarely exceed 500 to 600 miles in diameter. Individual storms may vary in size at different stages of their life. A storm generally increases its area as it moves along its track, so that a specific storm will cover the largest area during the later stages of its life. A case has occurred where a West Indian tropical revolving storm, retaining a vigorous circulation, has reached temperate latitudes. At this stage it was drawn into the warm sector of a depression and did not lose its identity in the circulation of the depression until just before reaching the British Isles.

WEATHER. Some tropical storms occur in which the wind never reaches

* Recent work suggests this statement should not be taken too literally. A few tropical storms are asymmetrical.

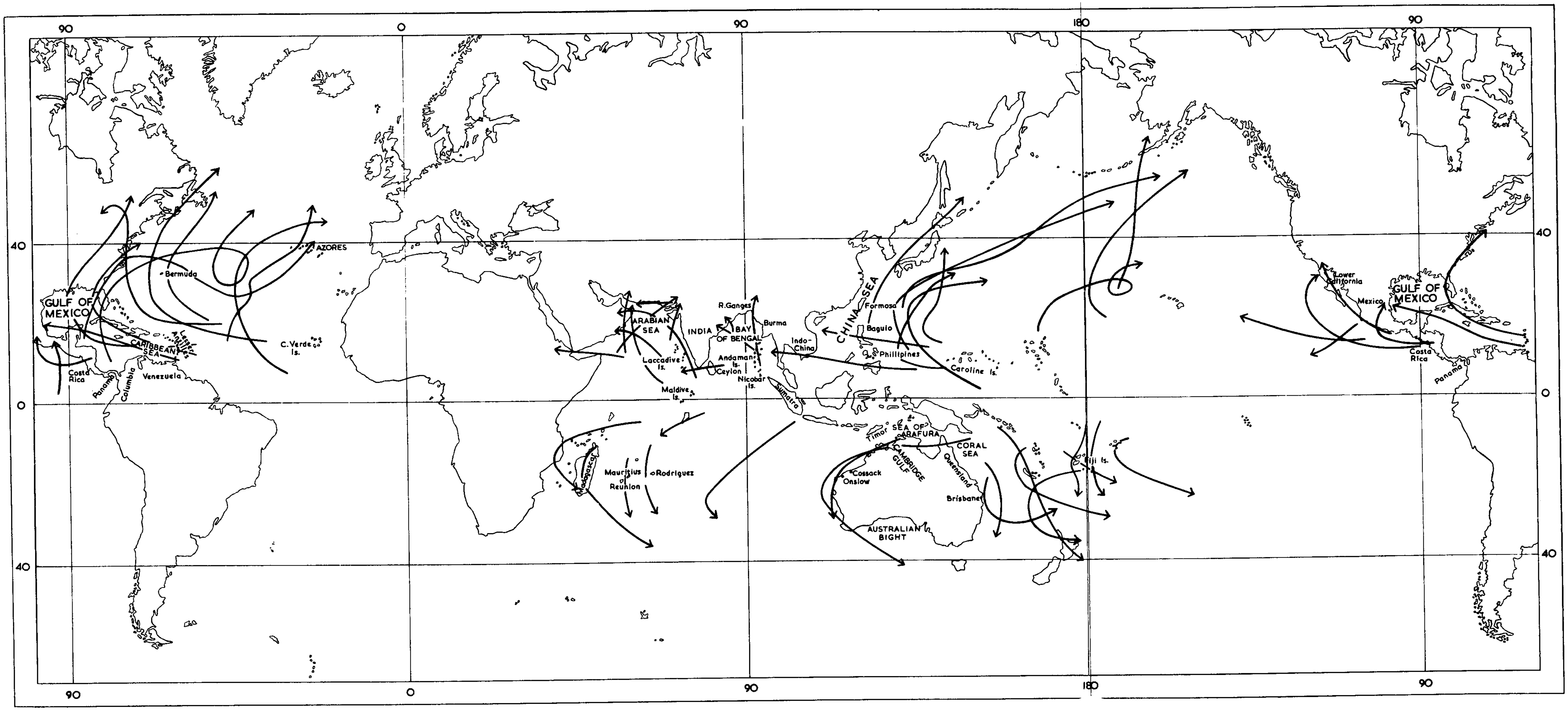

Fig. 11.1. Some tracks followed by tropical revolving storms

hurricane intensity. However, in the majority they do reach this force and are accompanied by torrential rain, sometimes with thunder and lightning, although the thunder may not be heard due to the roar of the wind and sea. The combined force of the hurricane winds, the high confused seas of the centre of the storm and the storm tides on the coasts, often result in extraordinary havoc. The outstanding features of a tropical revolving storm are the violent winds and intense rain squalls, except near the almost windless centre, vortex or 'eye' where there is usually only broken low cloud, also the mountainous seas and the accompanying noise effects.

WIND. Although winds of hurricane force are not unknown in the storms of temperate latitudes, there is an indescribable fury in the fully developed storm of the tropics. For a ship at sea (or sheltering in harbour) in the path of a tropical revolving storm, it is important to remember that after the vortex has passed the ship, winds of hurricane force begin to blow from the opposite direction to that from which the wind blew before the vortex arrived. The force of the wind varies greatly in different storms and in different parts of the same storm. Near the centre the winds are light and there is a nearly circular area of fair weather which is free from medium and high cloud and where the low cloud is often broken. This is called the 'eye' of the storm and usually occupies only a comparatively small area, up to 30 to 40 miles in diameter. High confused sea and swell usually prevails in the eye of the storm, constituting as great a danger to a ship as in the area of hurricane winds outside it. Away from the centre, winds of great violence occur, accompanied by even more violent squalls, with sudden shifts of direction. After the passage of the windless centre, the wind blows suddenly and furiously from a direction completely opposite to that from which it previously blew. The arrival of this new wind is heralded by an ominous roar which bursts upon the comparative stillness of the calm centre. In a violent storm the average wind strength might be Beaufort force 12 at 35 miles from the centre, force 11 at 50 miles from the centre, gradually decreasing outwards to about force 6 at a distance of 150 to 200 miles from the centre.

RAIN AND CLOUD. Heavy rain, sometimes with thunder and lightning, accompanies all tropical revolving storms. The rain usually begins at about 100 to 150 miles from the centre, but this varies in different storms. A rainfall of 20 in. within two days is not uncommon. The largest amount recorded in a typhoon was 46 in. in 24 hours at Baguio in the Phillippines. During the period of rainfall the sky is covered with dense Ns cloud and the visibility and air temperature decrease. In the central eye of the storm cloud diminishes, rain ceases, temperature rises and visibility improves. After the centre has passed, the sky again becomes overcast, rain begins with renewed violence and conditions return to those prevailing before the centre was reached. In the rear of the storm rainfall is not usually of long duration and the layer of dense cloud slowly gives way to Fc and Ci.

PRESSURE. In severe tropical revolving storms the pressure at the centre will be about 960 mb. However, much lower pressures may occur; one of the lowest observed at sea in a tropical storm was 914·6 mb on 5th November 1932 by the s.s. *Phemius* in the Caribbean Sea. There are not many observations of pressure at the centres of these storms, as ships normally try to avoid them. Over land the storms rapidly decrease in severity.

WAVES AND STORM TIDES. The advance of a tropical storm towards a coast is sometimes preceded by a storm tide. The water begins to rise on the coast, also because the storm loses kinetic energy in overcoming the friction of the ground. In spite of this decrease in intensity over land, cyclones of the Bay of Bengal occasionally cross India south of lat. 16°N and intensify again over the Arabian Sea. Similarly typhoons in the south China Sea sometimes cross the Kra Isthmus and renew their energies over the Bay of Bengal.

Frequency tables of tropical storm occurrences in different parts of the world are available, but in some areas the figures depend so much upon whether individual storms have chanced to cross a shipping route or an important island that they do not afford a basis for reliable comparison. Later in the chapter are shown tables giving the number of storms for each month over a number of years in various areas.

Fig. 11.1 gives some examples of storm tracks during the past half-century, and Fig. 11.2 illustrates some of the main characteristics of a tropical storm.

DIFFERENCE FROM TEMPERATE DEPRESSIONS. Tropical revolving storms differ from temperate depressions in other ways in addition to their smaller size and slower rate of travel. Since the Second World War much more has been learned about these storms. Aircraft have made reconnaissance flights which in some cases have penetrated the eye of the storm, and many photographs have been taken of the appearance of tropical revolving storms detected by radar on a PPI display; more recently satellite photographs have provided much useful information about them. These all show a number of characteristic belts of convective clouds, separated by narrower lanes of clear air, which spiral cyclonically towards the centre of the storm. Most of the squalliness and heavy rains of these storms is concentrated about these belts of convection, becoming more intense as the eye is approached; there is no evidence of any frontal structure resembling that found in a temperate latitude depression.

Cause and Formation

The causes of tropical revolving storm formation are not yet fully understood. However, a good deal has been found out about them by a study of the seasons, areas and synoptic situations in which they form, and the following conditions are now known to be favourable for their formation:

(*a*) They invariably form over the regions and during the seasons where and when the sea surface temperature is highest, i.e. usually in the western parts of the tropical oceans during their summer and early autumn.

(*b*) The existence of a deep, moist, unstable layer of equatorial air. Such layers are mostly found over the regions of highest sea temperatures.

(*c*) A small variation of wind with height (i.e. a small wind shear) throughout the lower troposphere.

(*d*) A previously existing weak cyclonic circulation (tropical depression). This is an essential condition for the development of a tropical revolving storm. It may appear as a closed circulation or as in (*e*) below.

(*e*) A region where there is a disturbance in the easterly trade wind circulation and where the air is moving cyclonically, e.g. on a curved path, keeping the low pressure on its left in the northern hemisphere.

There is little doubt that the reason why tropical storms only form over the regions of highest sea surface temperature, is because most of the energy of these ahead of the storm, a day or two before the storm itself arrives. At sea the winds in the storm field (*see* page 135) create waves whose height and length depend on the time they are influenced by the same wind direction and on the time its speed is maintained. The waves pass on through the storm field, and out ahead of it as swell. The water begins to rise along the coast when the storm is 300 to 500 miles distant and continues until the storm passes inland or curves and moves away. The rise of water may vary from 10 to 15 ft above the predicted tide level. Changes from the predicted maximum position of the storm tide may indicate changes in the track of the storm. Factors such as the configuration of the coastline and the distribution of islands will affect the storm tide.

MOVEMENT. Tropical revolving storms usually originate between approximately 8° and 20° latitude from the equator and travel to the westward. They cannot form in regions within about 5° of latitude from the equator because the earth's deflecting force, which is a function of latitude, becomes too small there for a cyclonic circulation to become established. When shallow areas of low pressure form near the equator, the air flow in such a region is indefinite and shows little dependence upon local values of pressure gradient. However, at latitudes more than 5° from the equator, the inflowing air becomes gradually deflected to the right in the northern hemisphere and to the left in the southern hemisphere, so that a circular vortex is formed, preventing the low pressure area from filling up and being destroyed.

The usual track of a tropical revolving storm follows a parabola around the oceanic 'permanent' high-pressure areas. Thus, after forming, the storm will move away to the westward on the equatorial side of the high-pressure system, tending polewards as it progresses. When the western side of the high-pressure system is reached, usually it will incline more and more polewards and then recurve to the eastward. The position where its westward movement changes to an eastward movement is known as the 'point of recurvature'. Thus in the northern hemisphere the general movement of tropical storms is roughly west, north-west, north and finally north-east, whereas in the southern hemisphere the corresponding directions are west, south-west, south and finally south-east. In both hemispheres the average positions of recurvature in the various oceans follow the movement of the overhead sun through the seasons in a manner closely resembling the seasonal migrations of the doldrums and subtropical high-pressure belts.

The rate of movement of tropical storm centres is generally less than 15 knots and therefore slower than that of temperate depressions. After recurving and acquiring an eastward movement, they begin to move rather faster and may reach 25 to 30 knots or more. However, a proportion of storms do not recurve, while the tracks of a few storms are most irregular and occasionally a small complete loop may be included in a track.* After reaching temperate latitudes, many storms tend to disappear by absorption into larger temperate depressions. Occasionally they may maintain their identity and become transformed into a temperate depression, acquiring a frontal system as they do so.

The storms gradually decrease in intensity as they move to higher latitudes, especially after recurving. They decrease considerably if they move inland, partly because there is then a wholly insufficient moisture supply available for providing the latent heat of condensation needed to energise the storm, and

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* *See Monthly Meteorological Charts of the Oceans*, published by the Meteorological Office, for selected tracks of tropical storms.

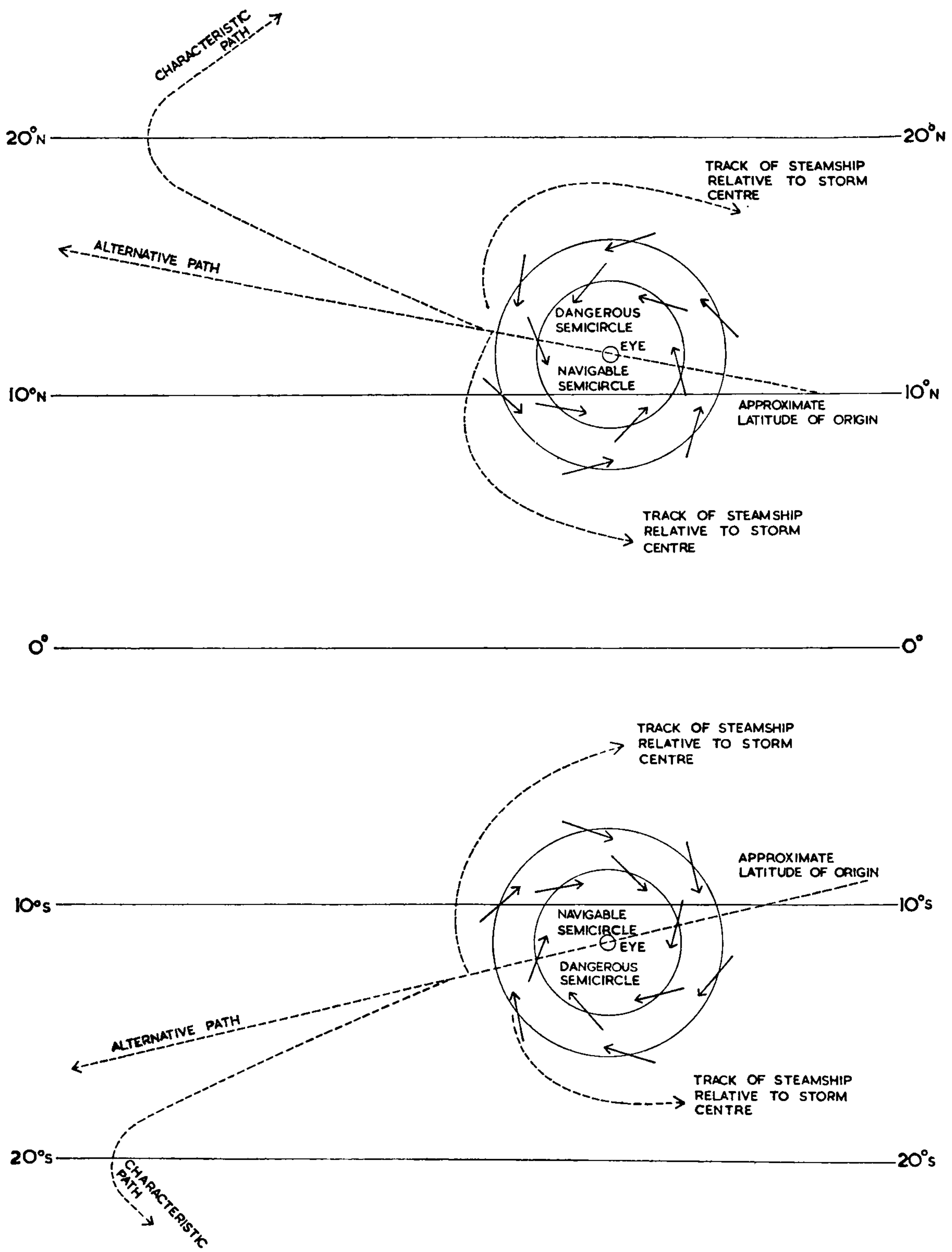

Fig. 11.2. Wind circulation and characteristic paths of tropical revolving storms

storms is derived from latent heat set free by condensation of water vapour in ascending currents of air. The largest supplies of water vapour are found in air which has moved over the warmest sea surfaces. In further support of this argument, tropical storms which leave the sea and travel over land soon decrease in intensity before they have travelled any great distance, and they have never been known to start developing over a large area of land.

It is also significant that tropical revolving storms are virtually unknown in the tropical parts of the South Atlantic and certain other low latitude regions, where the surface water is relatively cool. The fact that the equatorial trough (doldrums) does not penetrate into the South Atlantic may be related to the presence of the relatively cool sea surface there. It also implies that conditions for tropical revolving storm formation are not favourable because weak cyclonic circulations are unknown there, and hence condition (*d*) mentioned above is not satisfied.

These remarks give a general indication of the present theory of the formation of tropical revolving storms, about which there is still much to be learned.

Glossary of Terms Used

The following definitions are useful in connection with tropical storms.

VORTEX OR EYE OF THE STORM. The 'windless' central area within the ring of hurricane force winds of a tropical revolving storm, where the barometer is lowest.

CENTRE. That part of any cyclonic wind system where the barometer is lowest.

ANGLE OF INDRAFT. The angle which the direction of the wind makes with an isobar.

TRACK. The route along which the centre has travelled.

PATH. The term 'path' can indicate the route along which the centre has travelled or the route along which the centre is expected to travel.

POINT OF RECURVATURE. Whenever recurvature takes place this is the most westerly point reached by the centre when the storm is moving out of the tropics. Near this point the storm has usually moved round the western end of a subtropical anticyclone and is about to enter the circulation of the westerlies.

STORM FIELD. The regions covered for the time being by the winds forming the storm system.

RIGHT-HAND SEMICIRCLE. Looking along the path in the direction in which the storm is travelling, that half of the storm field which lies on the observer's right.

LEFT-HAND SEMICIRCLE. Looking along the path in the direction in which the storm is travelling, that half of the storm field which lies on the observer's left.

DANGEROUS SEMICIRCLE. This lies on the side of the path towards the usual direction of recurvature, i.e. the right semicircle in the northern hemisphere and the left semicircle in the southern hemisphere. This semicircle is so called because a ship caught in it may be blown towards the path over which the vortex will pass or the storm may recurve and the vortex pass over the ship. The advance quadrant of the dangerous semicircle is known as the dangerous quadrant.

NAVIGABLE SEMICIRCLE. The semicircle which lies on the side of the path away from the usual direction of recurvature, i.e. the left semicircle in the northern hemisphere and the right semicircle in the southern hemisphere.

Areas, Nomenclature and Seasons

The tropical revolving storm is known in various parts of the world by the local names of cyclone, hurricane, typhoon or willy-willy. Recent investigations into the synoptic situations of low latitudes indicates that there are more depressions in the tropics than was realised earlier, but that only in a small number does the wind attain hurricane violence (i.e. Beaufort force 12). The name cyclone originated in India because of the likeness between the spiral winds of the tropical storm and the coils of a snake. The word hurricane applied to the storms in the region of the West Indies comes from the vocabulary of an American Indian tribe and means 'big wind'. Typhoon is from the Chinese *tat* = great, *fung* = wind.

In addition to the names in the table below, there are various local names for them such as 'baguios' (Philippine Islands) and 'cordonazos' (west coast of North America). Whatever their name, they are all members of the same family and behave in accordance with certain broad laws and with equal ferocity. They occur chiefly in the late summer and early autumn months, and mostly over the western portions of the great oceans. The following table may be taken as a guide to their probable occurrence.

Table 11.1. Area, season and nomenclature of tropical revolving storms

Ocean	Season	Worst Months	Name
North Atlantic	June to November	September	Hurricane
Arabian Sea	Change of monsoon (but they have occurred in every month except February, March and August)	May and June; October and November	Cyclone
Bay of Bengal	All months except January, February and March	June and July; October and November	Cyclone
China Sea and W. North Pacific	July to November but no month is free	July and August	Typhoon
E. North Pacific	June to November	September	Hurricane
South Atlantic	No tropical revolving storm has ever been reported		
South Indian	November to April	January and February	Cyclone
W. South Pacific	December to April	January, February, March	Hurricane
N.W. Australia	December to April	January and February	Willy-willy

Brief Details and Behaviour in Various Areas

Tropical storms have been known to occur in the above-mentioned oceans outside the regions specified. The South Atlantic Ocean is the only ocean extending to the equator in which they have not been recorded.

It will be noted that the storms are seasonal and occur most frequently in late summer and early autumn. In the West Indies this well-known rhyme shows the probability of a storm occurring:

June—too soon
July—stand by
August—look out you must
September—remember
October—all over.

Actually hurricanes are by no means all over in October; in parts of the Caribbean and North Atlantic October can be one of the worst months. It was, for instance, on 5th November 1932, when the season should have been well past according to the rhyme, that the steamer *Phemius* sustained very severe damage in a West Indian hurricane and recorded a barometer reading of 914·6 mb, one of the lowest known.

West Indies and North Atlantic.* The total number of tropical storms

* A comprehensive history of individual storms which have been experienced in this area is given in *Hurricanes*, by I. R. Tannehill.

recorded each month over the 79 years 1886–1964 is as follows (U.S. Weather Bureau):

Jan.	Feb.	Mar.	Apr.	May	June	July	Aug.	Sept.	Oct.	Nov.	Dec.	Total	Yearly Average
0	1	1	0	10	41	46	142	216	146	29	4	636	8·1

Of the above, winds have reached hurricane force as follows:

Jan.	Feb.	Mar.	Apr.	May	June	July	Aug.	Sept.	Oct.	Nov.	Dec.	Total	Yearly Average
0	0	1	0	2	17	26	104	141	69	12	2	374	4·7

At the beginning and end of the hurricane season the area most favourable for hurricane development is in the doldrum region of the western Caribbean. In the middle of the season, however, the area of doldrums also covers the Cape Verde Islands and many hurricanes develop in that area.

The tracks of West Indian hurricanes are determined by the existing pressure distribution and chiefly by the position of the Azores anticyclone at the time. First moving to the westward, they seek to turn north at the earliest opportunity and tend to recurve into any existing trough of low pressure. An extension of the Azores anticyclone to the United States coast may prevent a hurricane from recurving. A storm thus prevented from following its normal track would continue in a westerly or north-westerly direction across the Caribbean Sea and Gulf of Mexico, dissipating inland over the southern States.

Most of the coasts of Venezuela, Columbia, Panama and Costa Rica are free from hurricanes, but rough seas and strong winds may be experienced in these regions as storms pass further northward.

Some of these storms proceed far up the American coast and past Bermuda; a few work their way with gradually diminishing strength across the temperate zone of the Atlantic towards Europe.

In October most of the storms originate east of the Lesser Antilles. The extension to the southward of the United States high pressure area causes them to be rather irregular in this month.

The average rate of progression of West Indian hurricanes is approximately 300 nautical miles per day. On the first leg of their path they move at about 11 knots. During recurvature, if the parabola of the path is an open one, their progress does not decrease much, but if the parabola is narrow the storm may become almost stationary. When obstructed by an area of high pressure, hurricanes may be stationary for two or three days. After recurvature their rate of progression increases to about 14 knots from May to August and to about 18 knots in September and October. If the hurricane continues into the temperate regions, the rate of progression increases to that of a depression of middle latitudes, i.e. between 20 and 30 knots.

The life of a hurricane while below the 30th parallel may be anything from one to 19 days but the average is about six days.

Arabian Sea. The total number of cyclones recorded in each month for the 66 years 1879–1944 is as follows (India Meteorological Department, *Storm Tracks in the Arabian Sea*, and *India Weather Review*):

Jan.	Feb.	Mar.	Apr.	May	June	July	Aug.	Sept.	Oct.	Nov.	Dec.	Total	Yearly Average
3	0	0	3	13	17	2	1	4	15	19	5	82	1·2

The cyclones chiefly occur in two distinct periods (April to July and September to January). The worst months, May–June and October–November, coincide with the change of the monsoon.

These cyclones are of two classes, those that are generated in the Arabian Sea and those that cross the peninsula from the Bay of Bengal. In the Arabian Sea itself they originate between the southern limit of the NE monsoon and the northern limit of the SW monsoon. Their exact birthplace thus varies with the seasonal movement of the monsoon, thus in May it lies between 5°N and 10°N, and 65°E and 75°E near the Maldive Islands. Later in the season it moves northward to the vicinity of the Laccadive Islands, between 9°N and 14°N, and 70°E and 75°E.

The Arabian Sea cyclones may follow a curved, straight or irregular path and their direction after recurvature may be anything between W and NE. Their tracks are very variable and it has been known for one to travel to the SW and for another to move ENE on the first stage of its journey. Their rate of advance changes during their own life and in the different seasons, but averages about 7 knots with a tendency to decrease somewhat during recurvature.

BAY OF BENGAL. The total number of tropical storms recorded in each month for the 61 years 1890–1950 is as follows (India Meteorological Department, *Memoirs*):

Jan.	Feb.	Mar.	Apr.	May	June	July	Aug.	Sept.	Oct.	Nov.	Dec.	Total	Yearly Average
9	2	4	25	48	76	113	129	132	119	92	44	793	11·7

Of the above, winds have reached storm force as follows:

Jan.	Feb.	Mar.	Apr.	May	June	July	Aug.	Sept.	Oct.	Nov.	Dec.	Total	Yearly Average
4	1	3	14	22	30	36	22	29	54	51	22	288	4·7

The worst months, May–July and October–November, coincide roughly with the change of the monsoon. The majority of these cyclones originate in the Bay itself, commencing during late April in the Sea of Andaman near the Nicobar Islands (6° to 9°N, 92° to 94°E). The zone moves with the sun and in June the southern half of the bay is clear and the cyclone birthplace is north of 16°N. In November it has moved south again to about 12°N.

There are also 'visiting storms' in the Bay of Bengal. These are typhoons which have moved westward from the North Pacific across Burma or Indo-China into the Indian Ocean. About one out of every five Bay of Bengal cyclones in September and October has this origin.

Cyclones in the Bay of Bengal can affect the whole of the adjacent coasts of India and Burma, and occasionally they can strike the NE coast of Ceylon. Their most common track is NW and thence N to the mouth of the Ganges, but there are instances of storms travelling in any direction between W and ENE.

Their rate of progression varies considerably in different storms. In their earlier stages they may be almost stationary and seldom advance at more than 4 knots. When fully formed they advance at about 12 knots. If they recurve, the rate of advance usually decreases whilst the recurve is in progress.

CHINA SEA AND WESTERN NORTH PACIFIC. The total number of typhoons occurring in the area bounded by latitudes 5°N and 30°N and longitudes 105°E and 150°E for the 70 years 1884–1953 is as follows (Royal Observatory, Hong Kong, *Technical Memoir No. 7*):

Jan.	Feb.	Mar.	Apr.	May	June	July	Aug.	Sept.	Oct.	Nov.	Dec.	Total	Yearly Average
18	10	10	22	69	108	263	305	310	213	148	65	1,541	22

Of 1,541 typhoons recorded about 40% occurred in August and September. There were less than 2½% in January, February and March which are the months in which least occur. The majority of these typhoons originate somewhere in the vast region between the Philippine Islands and 170°E, between lat. 5°N and 20°N. Although typhoons may originate anywhere in the above region, the most prolific source is near the Caroline Islands (5° to 10°N, 137° to 160°E). A few typhoons originate in the middle of the China Sea.

By far the greater number of the western North Pacific typhoons follow the accepted pattern of movement, namely, to the westward from their source and recurving to the NE on reaching higher latitudes. Others travel westward and dissipate over the Philippines or China coast. A few may find their way into the Bay of Bengal (*see* under Bay of Bengal). There is a case on record of a typhoon which approached the northern end of Taiwan in a WNW direction and then turned to the southward, passing through the Taiwan Straits from north to south. The course and speed of each typhoon is influenced by the existing general meteorological situation. Secondary typhoon centres, as violent as the primary, have been known to develop within the typhoon area. As with all tropical revolving storms, the speed of advance varies in each storm, but in general will be about 10 knots before recurvature and 20 knots after.

In addition to the general indications of the approach of a tropical revolving storm outlined later in this chapter, the direction of the wind in the seas adjacent to the northern Philippines may be a guide to the existence of a typhoon. In these waters a steady SW or NW wind should be suspected, since there is no season in the year when these winds predominate. (In this area the so-called SW monsoon more commonly blows from a direction between south and east.) Should the wind in the northern China Sea or Philippine Archipelago blow steadily from the SW during the months of June, July, August or September, the mariner may be fairly confident that there is a typhoon to the northward of him.

EASTERN NORTH PACIFIC. The total number of tropical storms recorded in each month for the 18 years 1947–64 is as follows (U.S. Weather Bureau):

Jan.	Feb.	Mar.	Apr.	May	June	July	Aug.	Sept.	Oct.	Nov.	Dec.	Total	Yearly Average
0	0	0	0	4	20	29	18	38	22	3	0	134	7·4

Of the above, winds have reached hurricane intensity as follows:

Jan.	Feb.	Mar.	Apr.	May	June	July	Aug.	Sept.	Oct.	Nov.	Dec.	Total	Yearly Average
0	0	0	0	2	7	12	5	15	13	2	0	56	3·1

Hurricanes in this area are also known locally as 'cordonazos', from a Mexican word meaning 'a lash of the whip'. They are confined to the south-eastern part of the eastern North Pacific and affect the west coast of the North American continent from Costa Rica to the middle of lower California.

The majority of these storms are born in the south-east portion of the area between 10°N and 30°N and eastward of 130°W to the coast, i.e. between the southern limit of the NE trade and the northern limit of the local SW monsoon.

These hurricanes usually move in a NW or WNW direction, roughly parallel to the coast; some recurve to the NE and disappear inland. Others, especially those which originate at some distance from the land, may move due W or even to WSW. Generally, in the early part of the season these storms occur at some distance from the coast, while from September onwards their tracks are much closer to it.

The rate of advance averages about 8 to 10 knots and the area covered by the storm field is usually of small diameter. When navigating in this region the mariner is at a disadvantage in that the usual timely indications of the approach of a storm may be lacking. If of small dimensions it may give little or no warning of its approach. This handicap is accentuated by the fact that between June and November the weather along the whole of the west coast of Mexico tends to be bad.

SOUTH PACIFIC OCEAN. The total number of hurricanes recorded each month for the 105 years 1819–1923 is as follows (Australian Commonwealth Meteorological Bureau, *Bulletin No.* 16, 1925):

Jan.	Feb.	Mar.	Apr.	May	June	July	Aug.	Sept.	Oct.	Nov.	Dec.	Total	Yearly Average
69	47	64	18	2	2	0	0	0	4	8	31	245	2·3

Hurricanes occur only in the western half of this area. Their place of origin is amongst the tropical island groups of the western South Pacific between 140°W and 160°E, and 5°S and 20°S. As the season progresses, the source region expands further westward, and by February hurricanes occasionally reach the coast of Queensland. Some also originate in the Coral Sea.

These hurricanes travel first in a SW direction, then curve to the southward, generally between 15°S and 25°S, and finally pass away to the SE.

The speed of advance varies: during its early life an average hurricane travels slowly but may reach 8 knots on the first leg; it slows down or even stops for a time at the point of curvature, after which it may increase to 15 knots or more.

SOUTH INDIAN OCEAN. The total number of cyclones recorded each month for the years shown is as follows (Netherlands Meteorological Service):

West of 80°E (1848–1947) (100 years)

Jan.	Feb.	Mar.	Apr.	May	June	July	Aug.	Sept.	Oct.	Nov.	Dec.	Total	Yearly Average
154	150	102	59	19	0	0	0	2	7	24	65	582	5·8

Between 80°E and 110°E (1848–1949) (102 years)

Jan.	Feb.	Mar.	Apr.	May	June	July	Aug.	Sept.	Oct.	Nov.	Dec.	Total	Yearly Average
22	22	38	28	8	3	2	0	0	2	17	15	157	1·5

Between 110°E and 125°E (1897–1949) (53 years)

Jan.	Feb.	Mar.	Apr.	May	June	July	Aug.	Sept.	Oct.	Nov.	Dec.	Total	Yearly Average
25	25	31	12	2	0	0	0	1	0	2	8	106	2·0

Cyclones in this ocean originate in the doldrum belt between the SE trades and the NW monsoon, which is an extension into the southern hemisphere of the NE monsoon. During the cyclone season (November to April) this doldrum belt extends in a NE–SW direction from Sumatra to northern Madagascar, and cyclones can develop anywhere in this belt.

The storms mostly follow the normal initial movement of the southern hemisphere cyclones to the WSW, curving gradually to the south and finally recurving to the SE. Recurvature of these storms normally takes place well to the east of Rodriguez in October and again in April and May, and between Rodriguez and Madagascar in November and December. In January, February and March they usually pass over Mauritius or Reunion during recurvature.

The rate of advance varies but usually averages about 8 knots, decreasing to about 6 knots during the recurve.

In the cyclones of the South Indian Ocean NE and E winds have been reported as blowing with a considerable component towards the centre, whilst they have on their southern side a strongly reinforced SE trade. This makes it difficult to tell when the trade winds form part of the storm area. A departure from the rules when navigating in these storm areas may therefore be necessary and is dealt with later in this chapter.

NORTH-WEST AUSTRALIA. The average number of tropical revolving storms is about one per year. (Season—December to April.)

The 'willy-willies' of this area originate in the Sea of Arafura near the island of Timor and normally travel from east to west, south of Timor, recurving to the SW. They may also originate in the vicinity of the Cambridge Gulf, travelling in a SW direction. Many of these storms reach the Australian coast in the vicinity of Cossack or Onslow, some pass southward along the west coast, whilst some of great intensity may even cross the Australian continent to regenerate as a lesser storm in the eastern portion of the Great Australian Bight.

Warning Signs

Apart from a definite, unusually steep and regular fall of the barometer, other indications, taken individually, are uncertain guides to the possible approach or development of a storm. Sea, wind direction and force, cloud, weather and barometer must be considered together, and when navigating in any area in which there is even a remote possibility of being visited by a tropical revolving storm, the mariner will do well to be constantly on the alert for any sign of a change in the weather.

Warning of the position, intensity and probable movements of a tropical storm may be received at any time by radio from a Meteorological Service or from another ship (*see Admiralty List of Radio Signals*, Vol. III). In most tropical storm areas the responsible Meteorological Services take considerable care to issue comprehensive warnings to shipping by radio when such a storm is known to be developing, and issue frequent bulletins concerning the storm's progress. In the hurricane season it is very important, therefore, that careful watch be kept aboard the ship for radio storm warnings. But as an official warning from a Meteorological Service is often based largely on observations from ships themselves, the shipmaster should be guided also by his own observations, bearing in mind that a Meteorological Service may not be in possession of sufficient information to issue even a general warning until some messages from one or more ships have been received. It is therefore very important that in the tropical storm season the master of every ship, if he suspects or knows of the existence of a tropical storm, should send a radio weather message to the nearest coastal radio station and to other ships as soon as possible, in accordance with the International Convention for the Safety of Life at Sea, and take action forthwith for the safety of his own ship. An intense storm of small diameter may develop with very little warning, and unceasing vigilance on the part of the master may well be the only safeguard against getting uncomfortably, or even dangerously, involved in such a storm. The following paragraphs are intended to help shipmasters in their assessment of the situation:

(*a*) If in a tropical storm area the barometer reading, corrected for height, latitude, temperature, index error and diurnal variation* is reading 3 mb or more below the mean pressure for the time of year, as shown in the *Admiralty Pilots* or *Meteorological Office Climatological Atlases*, the mariner should be on his guard. If the reading thus corrected is 5 mb or more below the normal, there can be little doubt that a tropical storm is in the vicinity, probably not more than 200 miles away. It is desirable to read the barometer at frequent intervals; the safest plan is to read it hourly in these areas during the cyclone season, even when conditions appear to be normal. *Any cessation of the diurnal range of pressure in the tropics should be regarded with suspicion.*

When a tropical storm passes fairly near to the ship there are usually three distinct phases in the fall of the barometer:

(i) A slow fall, with the diurnal variation still in evidence, usually occuring 500 to 120 miles from the storm's centre.

(ii) A more marked fall, during which the diurnal variation is almost completely masked. This usually occurs at from 120 to 60 miles from the centre. The barometer may be very unsteady throughout this phase.

(iii) A rapid fall occurring at from 60 to 10 miles from the centre.

(*b*) An appreciable change in the direction and strength of the wind during the cyclone season should be viewed with suspicion.

(*c*) These storms are frequently preceded by a day of unusual clearness and remarkable visibility. The atmosphere at such times is oppressive. These conditions are followed by extensive cirrus cloud, often V-shaped and pointing toward the storm centre. This cirrus shows no disposition to clear away at sunset but reflects lurid colourings then and also at sunrise. It later becomes reinforced by a thick layer of altostratus and ultimately by cumulus fractus and scud. The progress of the scud is marked by rain squalls of increasing frequency and violence. Rain is one of the most prominent features of a tropical revolving storm; in the outer portions it is intermittent and showery, whilst in the neighbourhood of the centre it falls in torrents. The rain area extends further in advance of the centre than in the rear.

(*d*) In the open sea, when there is no land between the ship and the storm centre, swell from the direction of the storm will probably give the first indication, since it travels at greater speed than the storm itself.

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* See the tables of diurnal variation in the relevant *Monthly Meteorological Charts*, or *Admiralty Pilots*; specimen tables (11.2 and 11.3) for the W. Pacific Ocean are given on page 150.

(*e*) The 3-cm type of radar normally fitted in merchant ships can, under normal meteorological conditions, be expected to detect moderate or heavy rain at its maximum range of about 30 miles. The 10-cm radar could be expected to detect such precipitation at considerably greater range. In either case sub-refraction would tend to decrease this range and super-refraction to increase it (*see* page 70). There is continuous heavy rain in a ring of radius at least 50 miles around the vortex of a tropical revolving storm. The utility of a ship's radar set in this connection is limited, because by the time that the rain comes within range of the radar set, the ship may well be already experiencing winds of force 9 or 10 with high seas, and though there may still be time, it would be imprudent to wait for this final and unmistakable sign of the proximity of the storm's centre before taking action. Very clear pictures of the centre of a tropical revolving storm have been seen on the PPI of a radar set. (*See* page 133.)

Action to be taken when in the vicinity of a Tropical Revolving Storm

With the preceding thoughts in mind, let us now assume that the master of a certain vessel considers that he is in the vicinity of a tropical revolving storm. His first care will be to place as much distance as possible between himself and the vortex and to warn other shipping of impending danger. It is realised that at such times masters are preoccupied with the safety of their own ships, but a concise weather report from a ship, either in code or plain language, will almost certainly result in timely advice being given to other ships who may be in the path and also to the inhabitants of island and coastal communities where life and property may be threatened. He should, therefore, lose no time in transmitting a priority message by radio, either in plain language or international code, addressed to the nearest coastal radio station and repeated CQ to all ships. This is required by Article 35 of the International Convention for Safety of Life at Sea. The text of such a message might read as follows:

TTT Storm. Appearances indicate approach of hurricane 1300 GMT July 10, 12° 22′N, 72° 36′W. Barometer corrected 994 mb, tendency down 6 mb. Wind NE force 8. Frequent rain squalls. Course 038°, 10 knots.

As long as the ship is under the influence of the storm similar messages should be transmitted at least every three hours. In such cases selected ships would send their messages to the appropriate shore authority in the international meteorological code, repeating it in plain language for the benefit of other ships.

In order that he may place his ship in a position of comparative safety the mariner should:

(*a*) Determine the bearing of the centre of the storm and endeavour to estimate its distance from the ship.
(*b*) Determine the semicircle in which the ship is situated.
(*c*) Plot the probable path.

It should be remembered that it is quite impossible to estimate the distance of the centre by the height of the barometer, or by its rate of fall, alone.

The bearing may be ascertained by the application of Buys Ballot's law, namely, face the wind and the low-pressure area will be on your right if you are in the northern hemisphere, and on your left if you are in the southern hemisphere. In northern latitudes the centre will bear about 12 points to the right at the beginning of the storm, i.e. when the barometer begins to fall. When the barometer has fallen 10 mb (3/10 in) the centre will bear about 10 points to the right, and when it has fallen 20 mb (6/10 in) the centre will be about 8 points to the right (to the left in the southern hemisphere). The nearer the observer is to the centre, the more nearly does the angular displacement approach 8 points, i.e. the direction of the wind more closely follows the isobars.

The distance of the centre from the ship will depend on so many factors that it is almost impossible to estimate without the aid of information from other sources. From the barometric pressure and force of the wind one can, however, arrive at certain broad conclusions. For instance, if the corrected barometer is 5 mb below the normal for the time of the year, the centre of the storm is probably not more than 200 miles away. At this distance the wind will probably have increased to force 6. If the wind is force 8, the centre is probably within 100 miles.

Determination of the semicircle in which the ship is situated depends on a true appreciation of whether the wind is veering, backing or remaining steady in direction. Unless the relative speeds and directions of the ship and the storm field are known (and, in the early stages, it is unlikely that they will be), the prudent shipmaster will at once heave-to and watch for a shift of wind, carefully watching the barometer at the same time. If he does not heave-to, the master must work out a relative motion problem to determine exactly the real wind shift. If the wind veers he is in the right-hand semicircle, if it backs he is in the left-hand semicircle, if it remains steady in direction he is in the path of the storm. The movement of the barometer will enable him to subdivide his semicircles into quadrants. The barometer falls before the trough and rises after it. The rules for determining the semicircle hold good in either hemisphere.

The path of the storm may be approximated by taking two such bearings of the wind direction with an interval of from two to three hours between them, provided that allowance is made for the ship's movement.

As it is so difficult to determine exactly the movement of the storm if the ship continues to make headway, the best and surest way of getting an accurate impression of the storm's movement is to stop the ship between the two bearings. It can normally be assumed that the storm is not travelling towards the equator and if in a lower latitude than 20° it is unlikely to have an easterly component.

A piece of tracing paper on which is drawn a diagram representing the average winds and tendency of the barometer in a tropical revolving storm (*see* Fig. 11.3) will be found useful for studying the behaviour of the wind in the storm field. One can lay off the ship's course and speed on a chart and manipulate the tracing paper as necessary to indicate the relative motion of the ship and the storm. Many examples which may be experienced according to the ship's position, course and speed relative to that of the storm can thus be illustrated, but it must be remembered that the forces and directions of the wind shown on the tracing paper are average or approximate.

Practical Rules for avoiding the Centre of a Tropical Revolving Storm

In whatever situation a ship may find herself, the matter of vital importance is to avoid passing within 50 miles or so of the centre of the storm; it is preferable to keep outside a radius of 200 miles or more, because at this distance the wind does not often exceed force 7 (and is generally not more than force 6) and

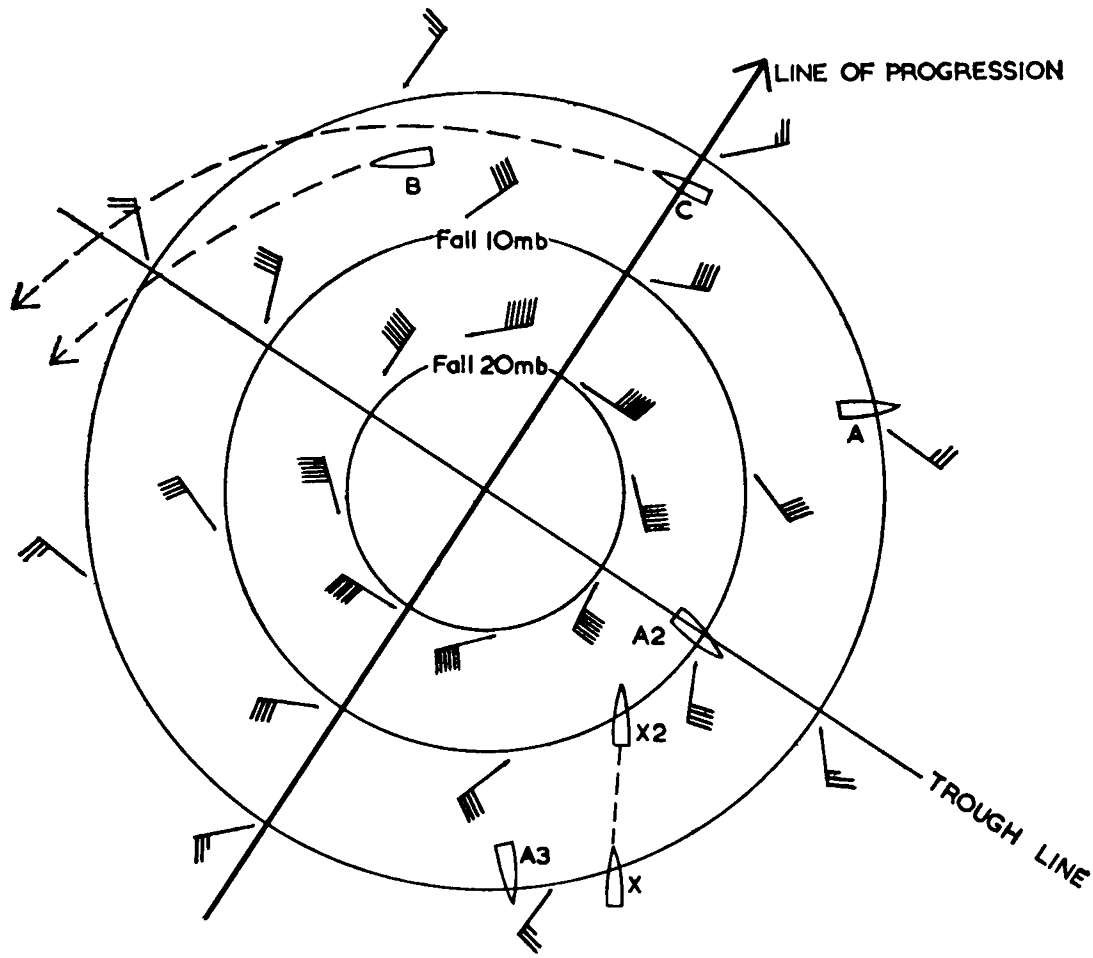


Fig. 11.3.

Rules for avoiding the centre of a tropical revolving storm (northern hemisphere)

(A description of these rules for northern and southern hemispheres is given on p. 146).

As the storm approaches. Long low swell, pressure falls more than 3 mb below normal value and diurnal variation ceases, sometimes remarkable visibility, extensive Ci cloud with lurid sunrises and sunsets. Wind and sea increase, pressure falls more rapidly, cloud thickens to As and scud. Squalls become more frequent and violent, torrential rain, sky covered with dense Cb cloud. Wind and sea continue to increase, visibility reduced to near zero.

After the storm has passed. Wind and sea decrease, slowly at first, rain lessens, cloud cover thins, pressure rises rapidly, then more slowly and resumes its diurnal variation and average height, swell decreases.

Vessel A. Is in the right-hand semicircle with a falling barometer. Heaves-to on the starboard tack or steams with the wind ahead or on the starboard bow and is thus heading away from the storm field. The barometer will cease to fall when she is in position **A2** and when at **A3** she will be able to resume her course again.

Vessel B. Is in the left-hand semicircle with a falling barometer. Runs with the wind on the starboard quarter whether steam or sail, hauls round as the wind backs and traces a course relative to the storm, as shown by the pecked line, with barometer rising slowly.

Vessel C. Is in the direct path of the storm and acts as vessel B.

Vessel X. Is overtaking the storm and converging on its centre. If the master obeys the rules and heaves-to, he will find the wind is veering and the barometer rising. He will thus know that he is in the rear quadrant of the right-hand semicircle and will, by keeping the wind on his starboard bow for a few hours in accordance with the 'rules', let the storm draw ahead before resuming his original course and getting astern of the storm. If, however, he does not heave-to, he will find that the wind is hauling to the left, relative to the ship, and with a falling barometer he might incorrectly assume that he is in the advance quadrant of the left-hand semicircle. If he acts on this assumption and, in accordance with the 'rules', puts the wind on his starboard quarter he may eventually find himself in the dangerous quadrant of the storm.

freedom of manoeuvre is maintained. If a ship has about 20 knots at her disposal and shapes a course that will take her most rapidly away from the storm before the wind has increased above the point at which her movement becomes restricted, it is seldom that she will come to any harm.

Sometimes a tropical storm moves so slowly that a vessel, if ahead of it, can easily outpace it or, if astern of it, can overtake it. Since, however, she is unlikely to feel seriously the effects of a storm so long as the barometer does not fall more than 5 mb (corrected for diurnal variation) below the normal, it is recommended that frequent readings should be made if the presence of a storm in the vicinity is suspected or known, and that the vessel should continue on her course until the barometer has fallen 5 mb, or the wind has increased to force 6 when the barometer has fallen at least 3 mb. If and when either of these events occurs, she should act as recommended in the following paragraphs until the barometer has risen above the limit just given and the wind has decreased below force 6. Should it be certain, however, that the vessel is behind the storm, or in the navigable semicircle, it will evidently be sufficient to alter course away from the centre.

In the northern hemisphere. If the wind is veering, the ship must be in the dangerous semicircle. A steam or other power-driven vessel should proceed with all available speed with the wind 1 to 4 points (depending upon her speed) on the starboard bow and should subsequently haul round to starboard as the wind veers. If a steamer has insufficient room to make much headway when in the dangerous semicircle, she should heave-to in the most comfortable position relative to the wind, preferably with the wind on her starboard bow so that she is heading away from the centre of the storm. A sailing vessel in these circumstances should heave-to on the starboard tack and haul round to starboard as the wind veers. If the wind remains steady in direction, or if it backs, so that the ship seems to be nearly in the path or in the navigable semicircle, a steam vessel should bring the wind well on the starboard quarter and proceed with all available speed, subsequently altering course to port as the wind backs. A sailing vessel under these circumstances should run with the wind on the starboard quarter, altering course to port as the wind backs. (Note: It is sometimes difficult to determine satisfactorily if indeed the ship is nearly in the path, particularly if in the dangerous semicircle, because the wind does not always behave according to rule.) Fig. 11.3 illustrates the rules to be followed in the northern hemisphere.

In the southern hemisphere. If the wind is backing, the ship must be in the dangerous semicircle. A steam or other power-driven vessel should proceed with all available speed with the wind 1 to 4 points (depending upon her speed) on the port bow and should subsequently haul round to port as the wind backs. If a steamer has insufficient room to make much headway when in the dangerous semicircle, she should heave-to in the most comfortable position relative to the wind, preferably with the wind on her port bow so that she is heading away from the centre of the storm. A sailing vessel in these circumstances should heave-to on the port tack and haul round to port as the wind backs. If the wind remains steady in direction, or if it veers, so that the ship seems to be nearly in the path or in the navigable semicircle respectively, a steam vessel should bring the wind well on the port quarter and proceed with all available speed, subsequently altering course to starboard as the wind veers. A sailing vessel in these circumstances should run with the wind on the port quarter, altering course to starboard as the wind veers. (Note: It is sometimes difficult to determine satisfactorily if indeed the ship is nearly in the path, particularly if in the dangerous semicircle, because the wind does not always behave according to rule.)

In either hemisphere, if there is insufficient room to run when in the navigable semicircle and it is not practicable to seek a safe and effective shelter before the storm begins to be felt, a steamer should heave-to in the most comfortable position relative to the wind and sea, bearing in mind the proximity of land, and a sailing vessel should heave-to on the port tack in the northern hemisphere and on the starboard tack in the southern hemisphere.

If a ship finds herself in the direct path of the storm and has no room to run into the navigable semicircle, she should consider, bearing in mind possible recurvature, whether she should endeavour to make her way into the 'dangerous' semicircle (where she may at least be better off than remaining in the direct path of the storm) and continue to steam to windward as fast as she can, so as to get as far as possible from the centre, or to heave-to.

In the South Indian Ocean on the south side of a cyclone a strong SE trade wind associated with a falling barometer may be experienced. It is difficult to know exactly when this wind becomes part of the circulation of an oncoming cyclone, though the falling barometer may give an indication. If in this area during the cyclone season the SE trade wind increases to a gale, the prudent shipmaster will stop his ship and obey the above rules. If the wind remains steady in direction and the barometer falls, he will assume that he is in the direct path of a tropical revolving storm. However, owing to the belt of intensified trade wind, it is not safe to assume this until the barometer has fallen 20 mb. An additional indication of whether the ship is in a tropical revolving storm field or merely in the belt of trades may sometimes be had from the behaviour of the clouds. If their movement is persistently more from the south than the surface SE wind, the wind being felt is probably part of a cyclonic circulation.

In order to be on guard for an erratic movement in the path of a tropical revolving storm, it is as well to plot a 'danger area' on the chart as an added precaution (*see* Fig. 11.4). From the reported position of the centre of the storm, lay off its track and the distance it is expected to progress in 24 hours. From the reported centre, lay off two lines 40° on either side of the track. With the centre of the storm as centre and the estimated progress in 24 hours as radius, describe an arc to cut the two lines on either side of the track. This will embrace the sector into which the storm centre may be expected to move within the next 24 hours. Although it is impossible to say with certainty how a particular storm will behave, the sector does at least provide a margin of safety and it can be reasonably expected that the storm's track will be somewhere inside the sector during the next 24 hours. In taking avoiding action, provided there is sufficient sea room, the mariner would do well to endeavour to get his ship outside of this sector as early as possible. If, after a few hours, the direction of the storm is found to have changed, another sector should be drawn with reference to the new estimated path of the storm and action taken to get out of that sector.

The most difficult situation is encountered when the ship finds herself at or near the point of curvature of the storm. This situation is shown in Figs. 11.5*a–d*, and it will readily be seen how necessary is a constant watch on the weather even when it has been decided in which semicircle the ship is situated.

Although the terms 'navigable semicircle' and 'navigable quadrant' have been used, very violent winds will be experienced in these areas and a vessel should clear them as soon as possible.

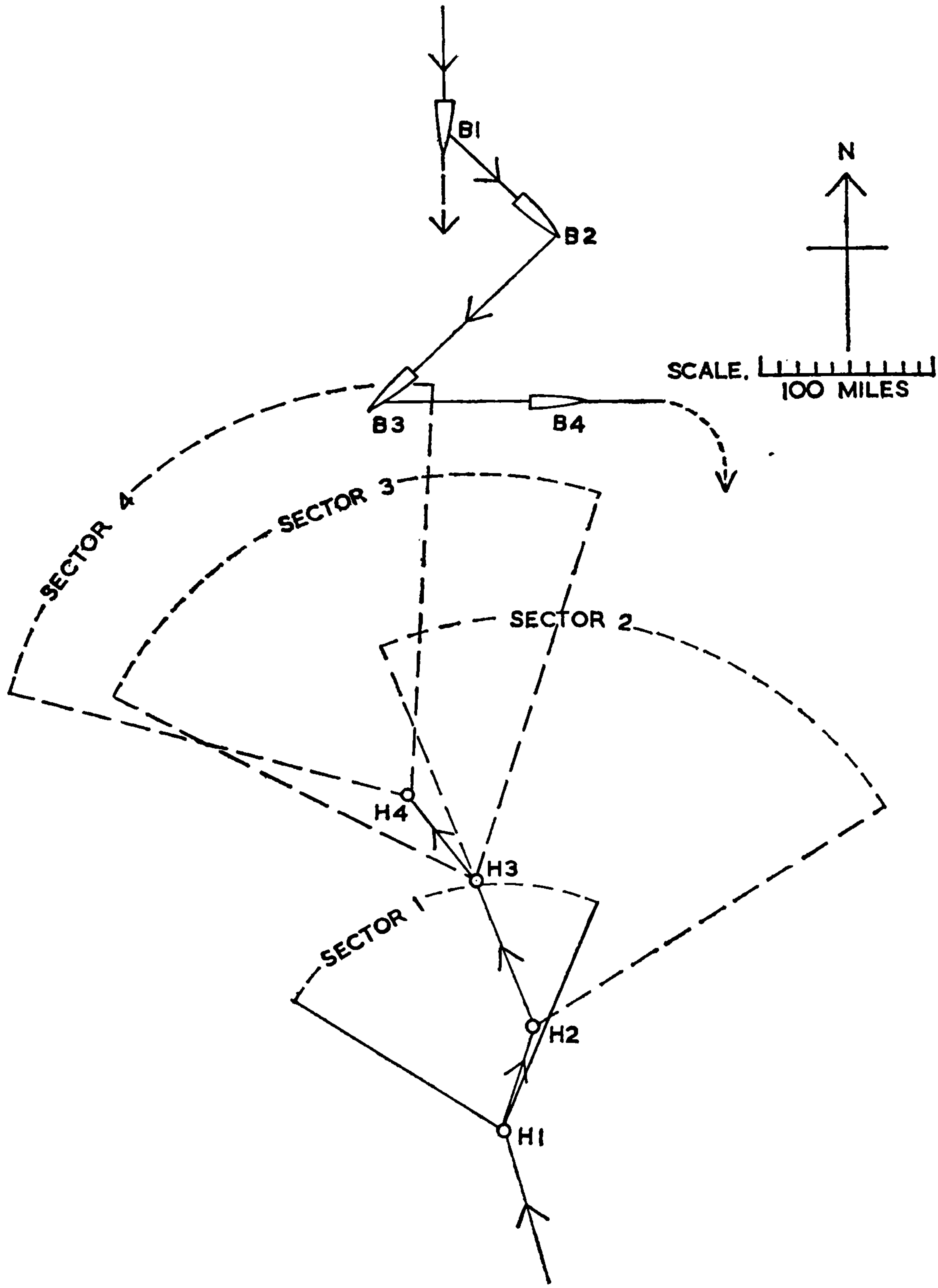


Fig. 11.4. Use of safety sector for keeping a ship clear of the track of a tropical revolving storm

A ship in position **B1** steaming south at 15 knots receives a report of a tropical revolving storm to the S of him with centre at **H1** and travelling about 340° at 6 knots. Sector 1 is drawn and course altered to SE to avoid it (the master considers that an alteration to SE is preferable to one to SW because, although it will place him on the eastern or dangerous side of the storm field, he will pass further away from it than if he altered to SW).

Six hours later, however, when the ship is at **B2** the centre of the storm is reported at **H2**, it having travelled in an 020° direction at about 10 knots. Sector 2 is now drawn on the assumption that this new path of the storm will be maintained. The master surmises that the storm has now recurved and therefore alters course to SW to get clear of the sector and eventually to pass astern of the advancing storm. Nine hours after this, when the ship is at **B3**, the storm centre is reported at **H3**, it having now resumed its original course. Sector 3 is now plotted and it is seen that the ship's present course will lead her into this new sector. The ship therefore makes a drastic alteration to E, this now being the quickest way of clearing the sector.

The storm centre is next reported at **H4**, six hours later, when the ship is at **B4**. Sector 4 is now drawn and it is seen that the ship is running clear, and with a watchful eye on further reports and his own weather, the master may soon be able to resume his original course.

It will be seen from the diagram that the safety sector is merely a rough-and-ready method of keeping clear of the storm field altogether. Its use depends on the reception of radio reports about the position of the storm centre and its progress, and its accuracy on the assumption that the storm will not alter course more than 40° without being detected. If no reports of the position and progress of the storm centre are received, it will be impossible to plot a sector and the mariner must be guided by his own observations and those of other ships in the vicinity, and by careful attention to the 'Practical rules for avoiding tropical storms'.

If in harbour, whether alongside or at anchor, the mariner should, in the cyclone season, be just as careful as at sea in watching his barometer and the weather and, if a tropical storm is threatened, in watching the shifting of the wind and estimating the movement of the storm relative to himself, so that he can take timely and seamanlike precautions well before the storm reaches his area. If the storm is known to be heading towards the harbour, it is often preferable to put to sea, provided there is plenty of sea room outside, rather than encounter the storm in harbour.

Riding out a tropical storm, the centre of which passes within 50 miles or so, in a harbour or anchorage, even if some shelter is offered, is an extremely unpleasant and hazardous experience, especially if there are other ships in company. The extreme violence and gustiness of the wind and its sudden shifts of direction involve great risk of the anchors dragging. The torrential rain and driving spray may impair visibility to such an extent as to make it very difficult to see if such is in fact the case.*

Discretion must, of course, be used. In the case of a low-powered or small vessel having insufficient warning to enable her to gain a reasonable distance from the storm or from the shore by putting to sea, it may be preferable to remain in a reasonably sheltered harbour. If a vessel of this type receives warning of an approaching storm when at sea, and there is considered to be insufficient time or sea room to avoid the dangerous part of the storm area, it may be advisable to seek shelter. In the China Sea, for example, there are so-called typhoon harbours which are listed in the *Admiralty Pilot*. In all cases, however, the mariner must use seamanship and initiative. It would, for instance, be imprudent to make for harbour and attempt to pick up hurricane moorings if the storm was already being felt. It would be equally imprudent to remain at anchor, even in a sheltered harbour, without raising steam as soon as the first warning of an approaching tropical storm is given.

* For some 10 hours during the worst of the hurricane, which devasted the town of Kingston, Jamaica, in August 1951, a British ship, having previously buoyed both her anchors with brightly painted drums and with plenty of cable out, maintained her anchorage by steaming so as to keep the buoys in sight ahead. The master reported that he averaged three-quarter speed on his engines throughout the period. (*The Marine Observer*, Vol. 23 (1953), page 112.)

TABLES* FOR THE DIURNAL VARIATION OF BAROMETRIC PRESSURE FROM 20°N TO 20°S BY 10° ZONES OF LATITUDE

Table 11.2. Correction to be applied to the observed pressure, for diurnal variation

Local Time		0	1	2	3	4	5	6	7	8	9	10	11
20°N—10°N	mb	−0·5	−0·1	+0·4	+0·7	+0·7	+0·5	+0·1	−0·4	−0·8	−1·2	−1·2	−1·0
	in	−0·015	−0·003	+0·012	+0·021	+0·021	+0·015	+0·003	−0·012	−0·024	−0·035	−0·035	−0·030
10°N—0°	mb	−0·6	−0·1	+0·4	+0·7	+0·8	+0·7	+0·2	−0·3	−0·9	−1·3	−1·4	−1·2
	in	−0·018	−0·003	+0·012	+0·021	+0·024	+0·021	+0·006	−0·009	−0·027	−0·038	−0·041	−0·035
0°—10°S	mb	−0·6	−0·2	+0·3	+0·6	+0·7	+0·5	+0·1	−0·5	−1·0	−1·4	−1·4	−1·2
	in	−0·018	−0·006	+0·009	+0·018	+0·021	+0·015	+0·003	−0·015	−0·030	−0·041	−0·041	−0·035
10°S—20°S	mb	−0·5	0·0	+0·4	+0·7	+0·7	+0·5	+0·1	−0·4	−0·9	−1·2	−1·2	−1·0
	in	−0·015	0·000	+0·012	+0·021	+0·021	+0·015	+0·003	−0·012	−0·027	−0·035	−0·035	−0·030

Local Time		12	13	14	15	16	17	18	19	20	21	22	23
20°N—10°N	mb	−0·5	+0·1	+0·7	+1·2	+1·3	+1·2	+0·9	+0·3	−0·2	−0·7	−0·9	−0·8
	in	−0·015	+0·003	+0·021	+0·035	+0·038	+0·035	+0·027	+0·009	−0·006	−0·021	−0·027	−0·024
10°N—0°	mb	−0·7	0·0	+0·7	+1·3	+1·5	+1·5	+1·1	+0·5	−0·2	−0·7	−1·0	−0·9
	in	−0·021	0·000	+0·021	+0·038	+0·044	+0·044	+0·032	+0·015	−0·006	−0·021	−0·030	−0·027
0°—10°S	mb	−0·6	+0·1	+0·8	+1·4	+1·6	+1·5	+1·1	+0·5	−0·1	−0·6	−0·9	−0·9
	in	−0·018	+0·003	+0·024	+0·041	+0·047	+0·044	+0·032	+0·015	−0·003	−0·018	−0·027	−0·027
10°S—20°S	mb	−0·5	+0·1	+0·8	+1·2	+1·3	+1·2	+0·8	+0·3	−0·3	−0·7	−0·9	−0·8
	in	−0·015	+0·003	+0·024	+0·035	+0·038	+0·035	+0·024	+0·009	−0·009	−0·021	−0·027	−0·024

Table 11.3. Average values of the barometric change in an hour, due to diurnal variation

Local Time		0–1	1–2	2–3	3–4	4–5	5–6	6–7	7–8	8–9	9–10	10–11	11–12
20°N—10°N	mb	−0·4	−0·5	−0·3	0·0	+0·2	+0·4	+0·5	+0·4	+0·4	0·0	−0·2	−0·5
	in	−0·012	−0·015	−0·009	0·000	+0·006	+0·012	+0·015	+0·012	+0·012	0·000	−0·006	−0·015
10°N—0°	mb	−0·5	−0·5	−0·3	−0·1	+0·1	+0·5	+0·5	+0·6	+0·4	+0·1	−0·2	−0·5
	in	−0·015	−0·015	−0·009	−0·003	+0·003	+0·015	+0·015	+0·018	+0·012	+0·003	−0·006	−0·015
0°—10°S	mb	−0·4	−0·5	−0·3	−0·1	+0·2	+0·4	+0·6	+0·5	+0·4	0·0	−0·2	−0·6
	in	−0·012	−0·015	−0·009	−0·003	+0·006	+0·012	+0·018	+0·015	+0·012	0·000	−0·006	−0·018
10°S—20°S	mb	−0·5	−0·4	−0·3	0·0	+0·2	+0·4	+0·5	+0·5	+0·3	0·0	−0·2	−0·5
	in	−0·015	−0·012	−0·009	0·000	+0·006	+0·012	+0·015	+0·015	+0·009	0·000	−0·006	−0·015

Local Time		12–13	13–14	14–15	15–16	16–17	17–18	18–19	19–20	20–21	21–22	22–23	23–0
20°N—10°N	mb	−0·6	−0·6	−0·5	−0·1	+0·1	+0·3	+0·6	+0·5	+0·5	+0·2	−0·1	−0·3
	in	−0·018	−0·018	−0·015	−0·003	+0·003	+0·009	+0·018	+0·015	+0·015	+0·006	−0·003	−0·009
10°N—0°	mb	−0·7	−0·7	−0·6	−0·2	0·0	+0·4	+0·6	+0·7	+0·5	+0·3	−0·1	−0·3
	in	−0·021	−0·021	−0·018	−0·006	0·000	+0·012	+0·018	+0·021	+0·015	+0·009	−0·003	−0·009
0°—10°S	mb	−0·7	−0·7	−0·6	−0·2	+0·1	+0·4	+0·6	+0·6	+0·5	+0·3	0·0	−0·3
	in	−0·021	−0·021	−0·018	−0·006	+0·003	+0·012	+0·018	+0·018	+0·015	+0·009	0·000	−0·009
10°S—20°S	mb	−0·6	−0·7	−0·4	−0·1	+0·1	+0·4	+0·5	+0·6	+0·4	+0·2	−0·1	−0·3
	in	−0·018	−0·021	−0·012	−0·003	+0·003	+0·012	+0·015	+0·018	+0·012	+0·006	−0·003	−0·009

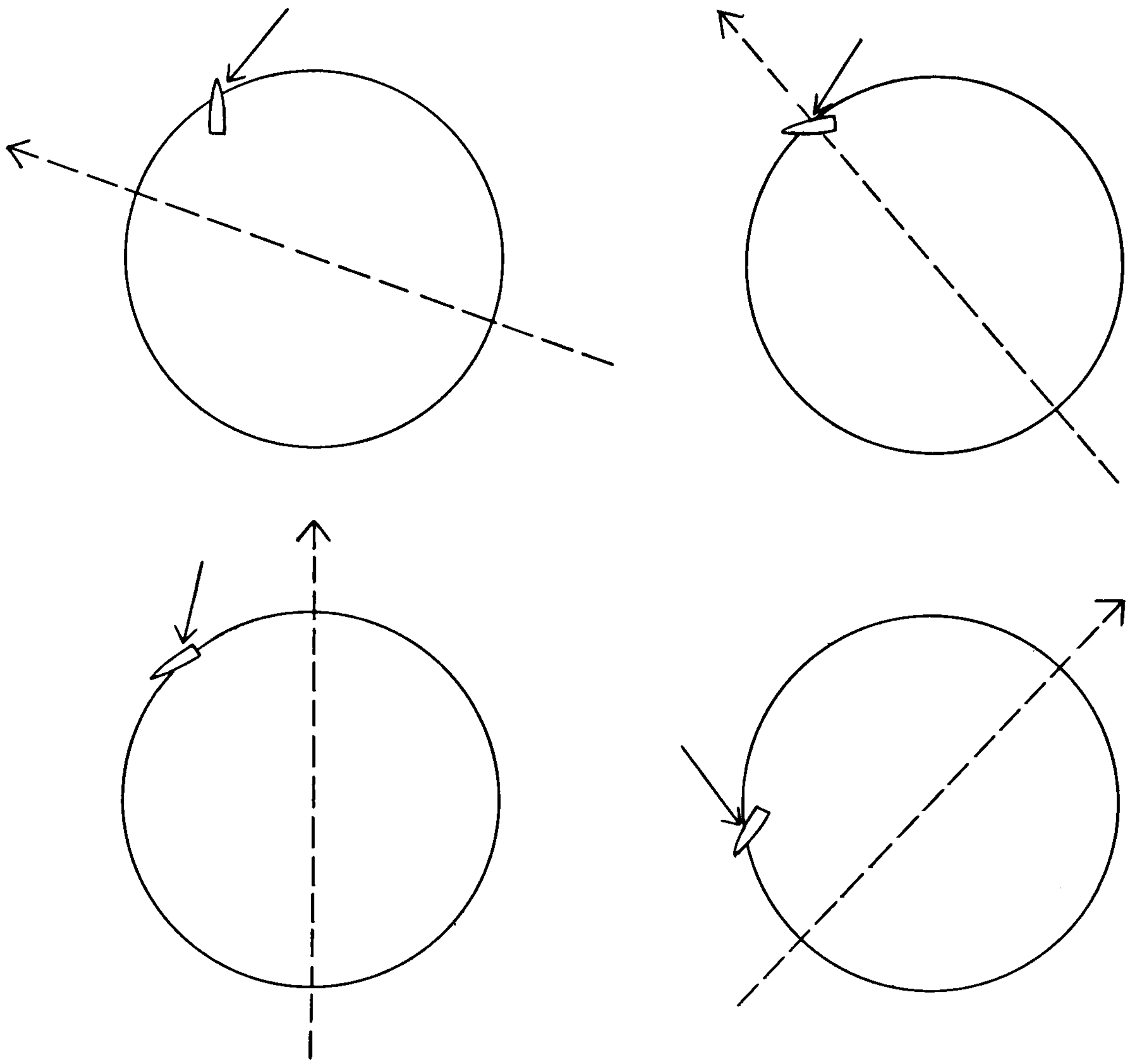
For the zones of latitude 20°N–10°N, 10°N–0°, 0°–10°S and 10°S–20°S, Table 11.2 gives the correction to be applied to the barometric pressure at each hour, to allow for the diurnal variation. Table 11.3 gives the average rise or fall during each hour, caused by this diurnal change.

The tables are based on observations made in British ships, at the hours 0000, 0400, 0800, 1200, 1600, 2000, local time, and forwarded to the Marine Division of the Meteorological Office. They extend over the years 1919–38.

In the tropics, should the barometer, after correction for diurnal variation (Table 11.2) be as much as 3 mb (approximately 0·1 in) below the monthly normal for the locality, as shown on the meteorological charts, the mariner should be on the alert, as there is a distinct possibility that a tropical storm has formed, or is forming. A comparison of subsequent hourly changes in his barometer with the corresponding figures in Table 11.3 will show whether these changes indicate a real further fall in pressure and its amount.

Note: When entering a barometric pressure in the log, or when including it in a wireless weather report, the correction for diurnal variation must not be applied.

* These tables apply only in the W. Pacific Ocean; tables for the other oceans can be obtained from the relevant *Monthly Meteorological Charts*. (*See* footnote on p. 142).



Figs. 11.5a–d. Movements of a ship in a tropical revolving storm at the point of recurvature

This illustrates the necessity of watching the wind direction carefully when in the vicinity of a tropical revolving storm. It is here assumed that a steamer is in the path of an advancing storm in the northern hemisphere and that, unknown to the meteorological authorities or the master of the ship, the storm has reached the point of recurvature. From a succession of observations made when the ship was hove-to, head on to wind and sea, it was found that the wind was shifting to the right and it was, accordingly, assumed that the vessel was in the right-hand semicircle of a storm moving to the NW.

During the interval between Fig. 11.5*a* and 11.5*b* the wind was observed to change very slightly, a little to the right at first, then back to the previous direction again but gaining in force. From this observation, it was correctly assumed that the vessel was now in, or nearly in, the direct path of the storm and accordingly course was altered to bring the wind on the starboard quarter as in Fig. 11.5*b*.

From that time onwards the wind was observed to shift progressively to the left as in Fig. 11.5*c*, suggesting that the ship was now in the navigable semicircle. The wind continued to shift to the left more rapidly, with the barometer rising, which showed that the storm was clearing to the eastward (*see* Fig. 11.5*d*).

Had the manoeuvre shown in Fig. 11.5*a* been continued without paying careful and frequent attention to the shifts of wind being experienced, the ship might have been badly involved in the direct path and actually in the centre of the storm and suffered severely in consequence.

PART IV. WEATHER FORECASTING

CHAPTER 12

THE WEATHER MAP

General Remarks

In this chapter the main considerations underlying the preparation of the weather map are described, together with a brief account of the way in which the meteorologist uses weather maps in order to identify air masses and fronts and to prepare weather analyses and forecasts.

The weather map is the basic tool of the forecaster. On it are plotted observations of pressure, temperature, wind and other weather elements received from a large number of stations. These observing stations form a close network over the land. At sea, observations are provided voluntarily by merchant ships and from ocean weather stations. It is common practice for all these observations to be made at the same time (GMT). The map therefore gives a bird's-eye view, or synopsis, of the weather over a large area at a particular time; hence the term synoptic chart.

The advantage of a synoptic chart for studying the weather was apparent long before the invention of radio. Admiral FitzRoy began collecting reports daily by telegraph in London in 1860 from which weather maps were drawn, and he shortly afterwards organised a system of storm warnings. In 1868, Alexander Buchan, using observations taken from ships' meteorological log-books, drew up a series of synoptic charts to show the travel of depressions over the Atlantic. An example of these charts is shown in Fig. 12.1. Later the Meteorological Council published a series of synoptic charts of the Atlantic for the year 1882–83. The observations on these charts were extracted from ships' logs long after the actual date, and hence the charts were only of value for investigation (Fig. 12.2). It is interesting to compare these with the modern synoptic charts as drawn up by selected ships making use of the Atlantic Weather Bulletin, which are shown at Figs. 12.3 and 12.4.

From the synoptic viewpoint, some inherent differences between ship and shore stations should be considered. Land stations form a fixed network whose density and distribution may be varied at will, according to requirements. If an additional reporting station is required in a certain locality, there is usually no great difficulty in establishing one, except in isolated areas. At sea, however, there are few reporting stations in fixed positions. The forecaster must rely upon ships' reports and hope to have a good enough selection to complete his chart over the ocean. On some occasions an additional ship's report in a key position makes a great difference in drawing up a synoptic chart. A case in point arose at 0600 GMT on 12th September 1951, when the solitary observation from s.s. *Ruahine* in position 42·6°N, 27·7°W made it possible for the Central Forecasting Office at Dunstable to determine the position of the secondary centre of low pressure shown in the lower left-hand corner of Fig. 12.3 with considerable accuracy. The chart for 0600 GMT on the 13th shows that 24 hours later this secondary centre had become an important feature of the weather over the British Isles. (*See* Fig. 12.4.)

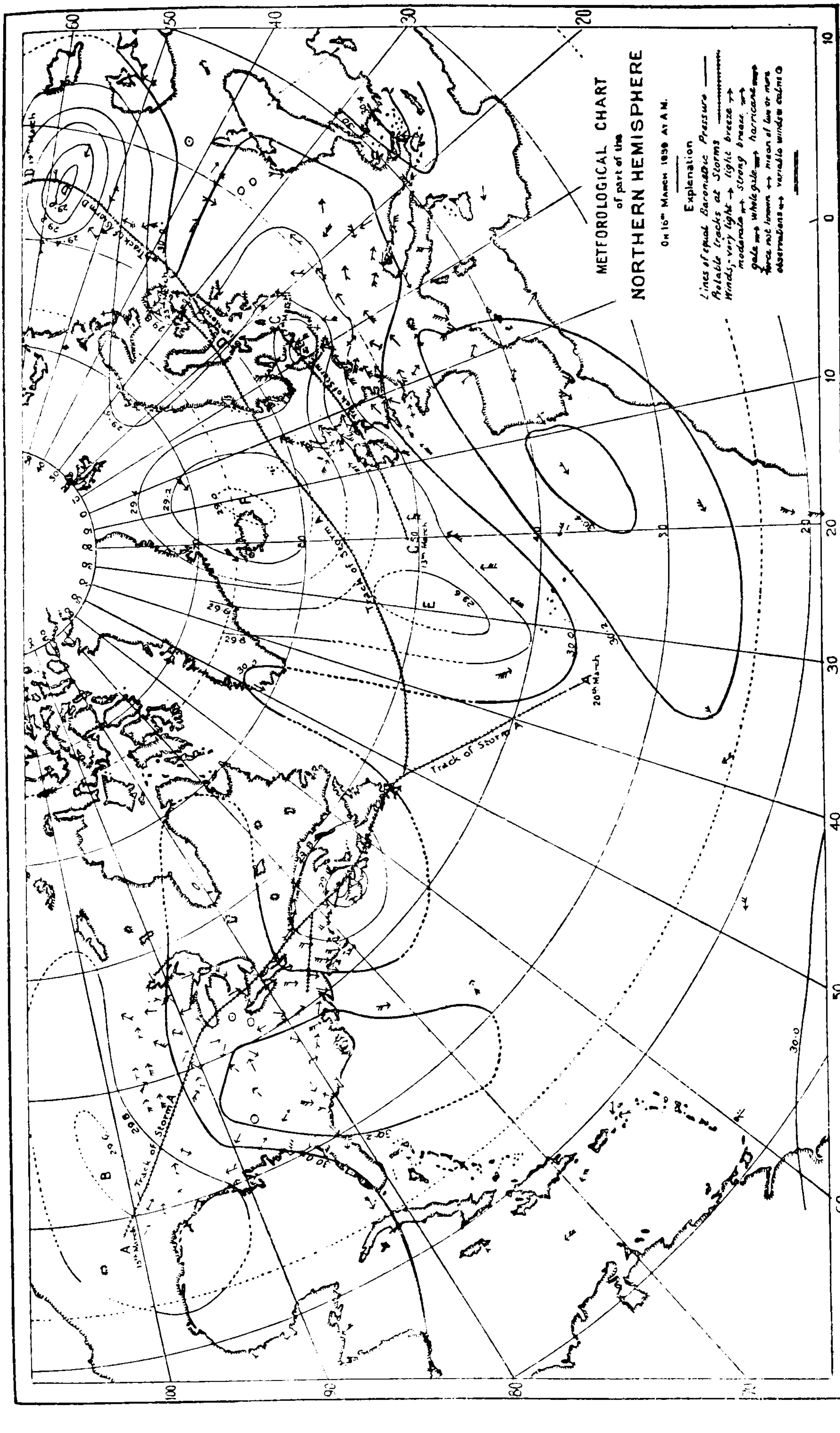


Fig. 12.1. Weather map of North Atlantic, 16th March 1859

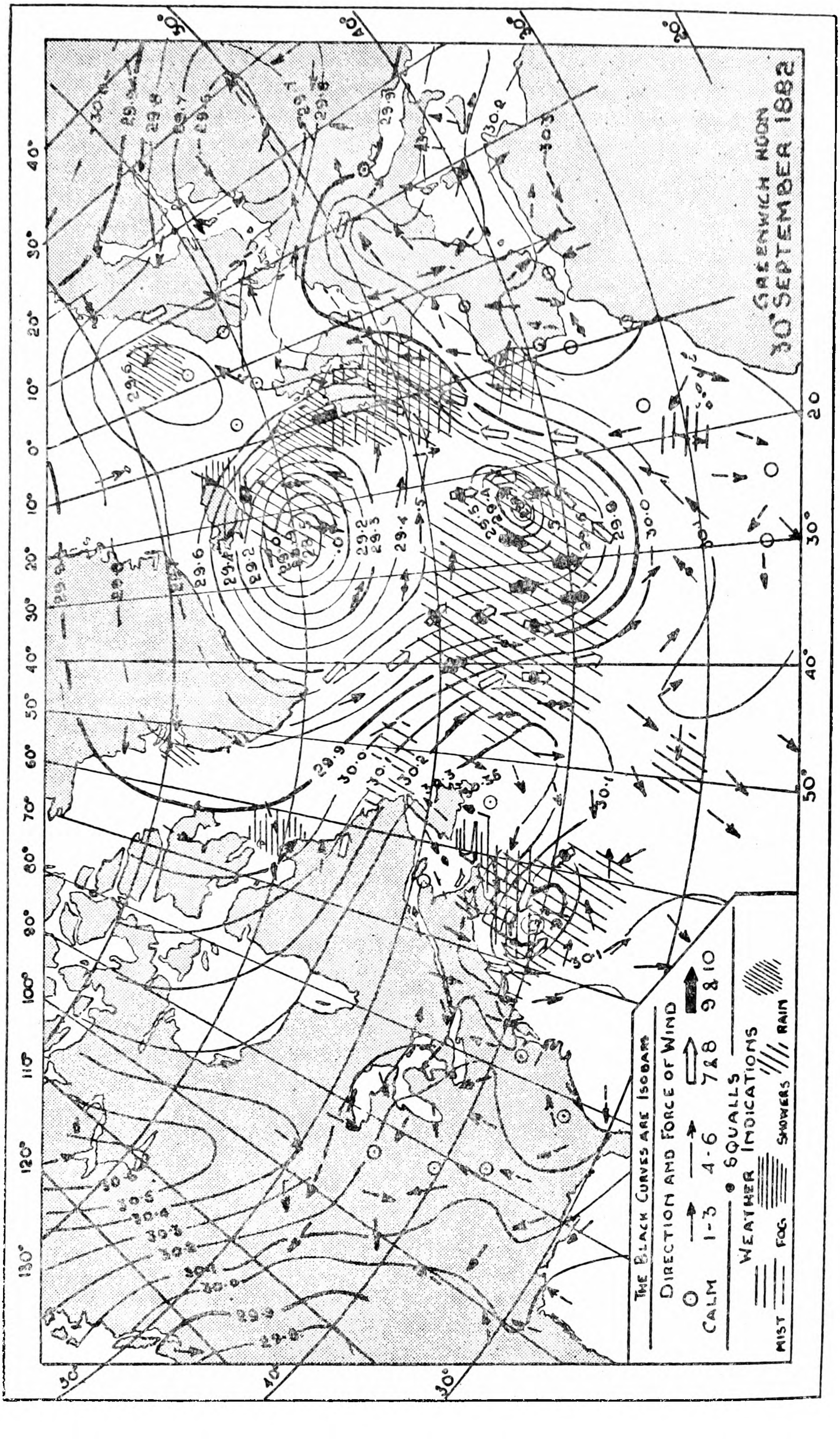


Fig. 12.2. Weather map of North Atlantic, 30th September 1882

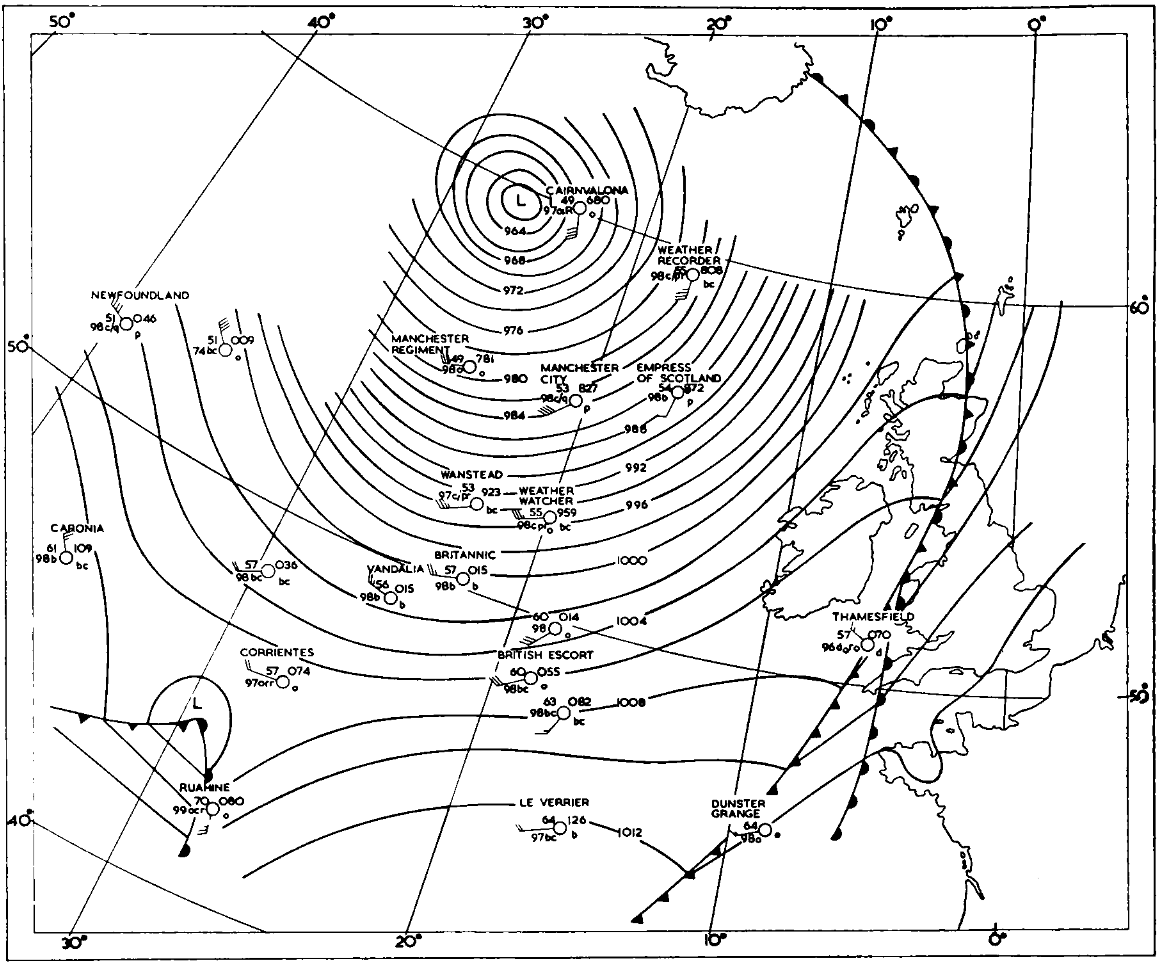


Fig. 12.3. Synoptic chart of North Atlantic, 0600 GMT, 12th September 1951

Representative Observations

The assumption that observations made at different stations are comparable and representative forms the basis of synoptic meteorology. To ensure comparable observations, both instrumental equipment and observing technique must be standardised. This is covered by the official instructions issued by the various national meteorological services, which are co-ordinated by the World Meteorological Organization. The aim is that different observers making a weather observation at the same place and time should get identical results, despite minor differences of aptitude and training. The fact that observations are comparable, as far as the forecaster is concerned, also means that each observation is representative, i.e. that it is free from any local influence which might prevent its being regarded as a fair sample of the weather in its neighbourhood.

A barometer reading aboard a ship, reduced to standard latitude and mean sea level, is representative except when affected by the movement of the ship ('pumping') and the suction effect of the wind on the instrument, due to its location in a partially enclosed space in the ship (e.g. the chart room).

Air temperature and humidity are usually measured by wet and dry-bulb thermometers, either in a portable screen or fitted as an aspirated psychrometer. Dry-bulb temperatures at sea are representative when adequate precautions are taken to guard against heat from the ship. For this reason they can be more useful to the forecaster for air-mass analysis than are temperature readings on land. Land readings, especially those from inland stations, are influenced by solar heating, nocturnal cooling and effect of altitude, which severely limit their

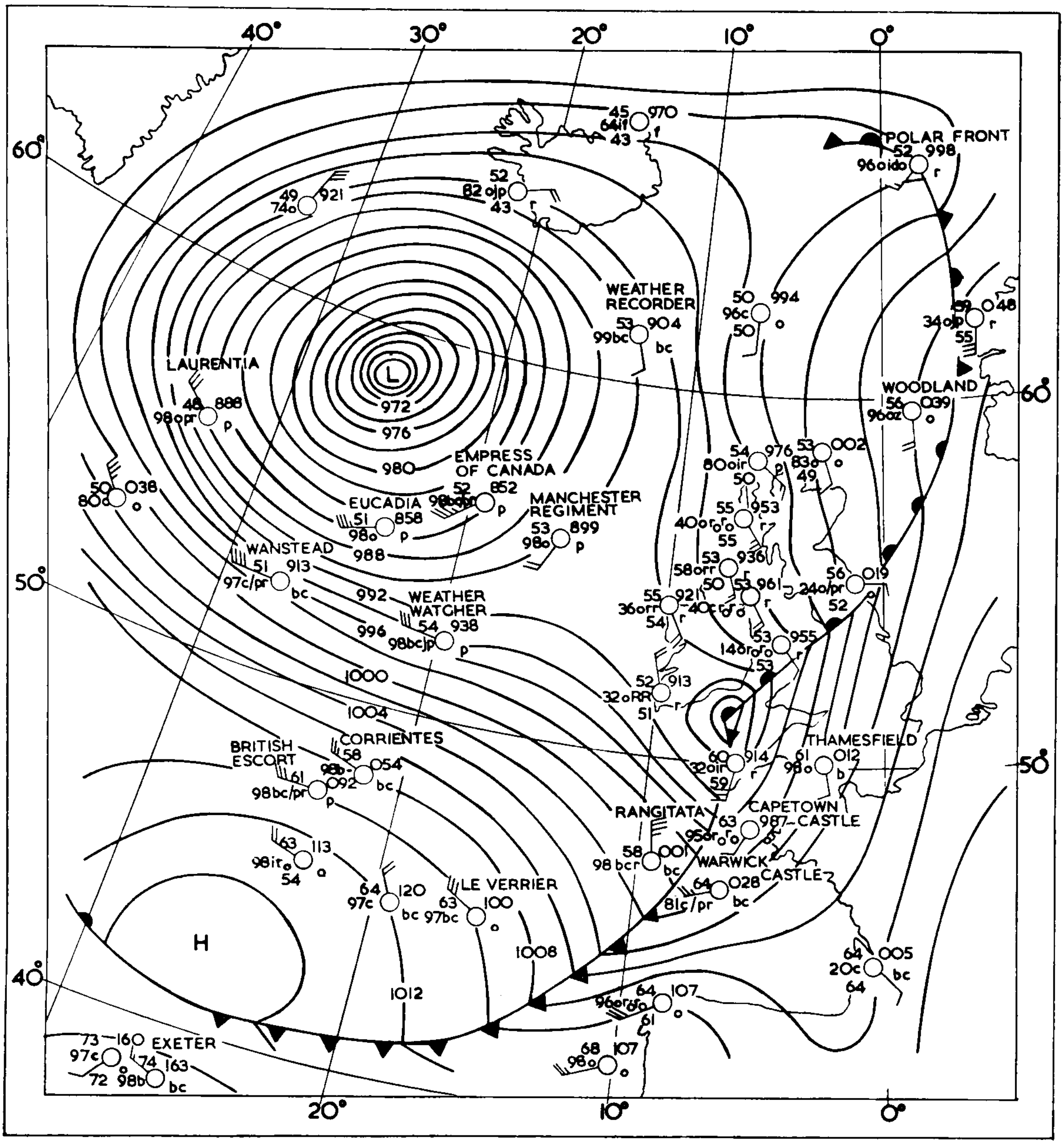


Fig. 12.4. Synoptic chart of North Atlantic, 0600 GMT, 13th September 1951

use in this respect. Precautions to be taken aboard ship to ensure a representative value of dry and wet-bulb temperatures have been emphasised in the *Marine Observer's Handbook.* Humidity readings at sea are normally less representative than at land stations, partly because of observational difficulties and partly because the readings vary through a smaller range, being usually nearer to saturation level than on land; hence experience and caution are required in using them for synoptic analysis.

The observation of sea surface temperature by means of a sample collected in a bucket affords an excellent example of the difficulties which must be overcome before a representative value of a meteorological element at sea can be obtained. The meteorologist is interested in the temperature of the surface, hence the sample which his bucket collects should preferably be from an undisturbed part of the surface and also not be contaminated by water discharged from the ship. Cooling or heating processes are likely to start altering the temperature of the water in the bucket immediately it has been pulled up. The cooling effect may be appreciable if the air temperature is much lower than the sea temperature and a breeze is blowing. The temperature has to be read, therefore, as soon as possible after drawing the water (within 30 seconds of hauling on deck). In some ships the temperature of the water in the engine intake, some 15 to 20 ft below the water line, is recorded instead of actual surface temperature. In moderate and strong winds and under cloudy skies there is often no difference between the value of sea temperature as measured from a sample collected in a bucket and from the sea water flowing through the engine intake; in other circumstances, such as light winds, calm seas and clear skies, there may be an appreciable temperature gradient in the layers nearest the surface; the engine intake temperature is not then representative of the sea-surface temperature.

On land, representative values of wind are obtained by erecting an anemometer at a standard height (generally 10 metres) on open, level ground, free from obstructions. The use of an anemometer at sea involves the major difficulty of finding a satisfactory exposure. Aboard merchant ships, wind speed and direction are normally estimated directly from the appearance of the sea surface. The method is independent of the motion of the ship and, in the open sea, the result is perhaps more representative than an anemometer reading, provided the wind is not changing rapidly.

Time Factor

Once the observations have been plotted on the synoptic chart, the forecaster can see how the weather is varying from one place to another. But he is also interested in the change of the weather with time, and some of the observations themselves give information about this. An example that comes to mind is the barometric tendency, or change of pressure during the past three hours. Barometric tendency, as received from the ship, is not representative until it has been corrected to allow for the course and speed of the ship. The past weather and the present weather, taken together, give a comparison of weather over a short period of time. In addition, the idea of sequence is introduced in the present weather code by the use of descriptive terms, such as 'intermittent' and 'continuous', and in the codes for reporting the state of sky by descriptions such as the following:

'Stratocumulus formed by the spreading out of Cumulus.'

'Cirrus (often hook-shaped) gradually spreading over the sky and usually thickening as a whole.'

In synoptic meteorology it is important that observations are taken at fixed times (synoptic hours), the charts drawn for each of these times forming a sequence whereby the 'time changes' of any element can be estimated. The shorter the time interval between the charts, the nearer the sequence of weather shown thereon will approach reality. As an illustration, think of some game the progress of which can only be watched from a series of photographs taken at equal time intervals. Intervals of five minutes would be far too wide to give any idea of the game, though one might see if the play was nearly always in favour of one side. If photographs were taken at cinematograph speed an excellent idea of the game would be afforded. The weather is a gigantic game whose state of play is only revealed to the forecaster at the synoptic hours. At present these are 0000, 0600, 1200, 1800 GMT, together with supplementary times 0300, 0900, 1500 and 2100 GMT. These supplementary times are used at land stations but not as a rule at sea. It is therefore important that ships' observations, being available only at six-hour intervals, should be of such quantity and quality as to enable the best possible charts to be drawn at the main synoptic hours.

Summarizing the synoptic method, we see that it not only allows comparison of weather from place to place at a fixed time (the synoptic hour), but by a sequence of charts permits 'time variation' of weather at each place to be estimated. The analysis of synoptic charts involves a combination of these two aspects.

Representation of the Data

To show each individual observation clearly on working charts as used by national Meteorological Services, some abbreviated notation is necessary. Fig. 12.5 shows some of the symbols used by international agreement. It will be noticed how these symbols are suggestive of the element represented. In order to avoid confusion, the symbols are grouped in a special way around the position of the station. Fig. 12.6 shows this station model together with some examples. The reader should note how concisely and conveniently this system expresses a mass of detail. The procedure is similar for the mariner wishing to plot his own synoptic chart while he is at sea, from the reports in the Atlantic Weather Bulletin and similar bulletins; though it is simplified by using the Beaufort letter abbreviations instead of symbols, details of which are given in Met. O. 509, Sixth Edition (Table XXXI), *Ships' Code and Decode Book*. In both these cases the method of plotting of the weather charts aims at portraying the information from a large number of coded messages in a manner in which it is as easily grasped as possible by the professional forecaster, or by the mariner for whom it is intended. When the plotting is completed the drawing up and analysis of the chart can be undertaken.

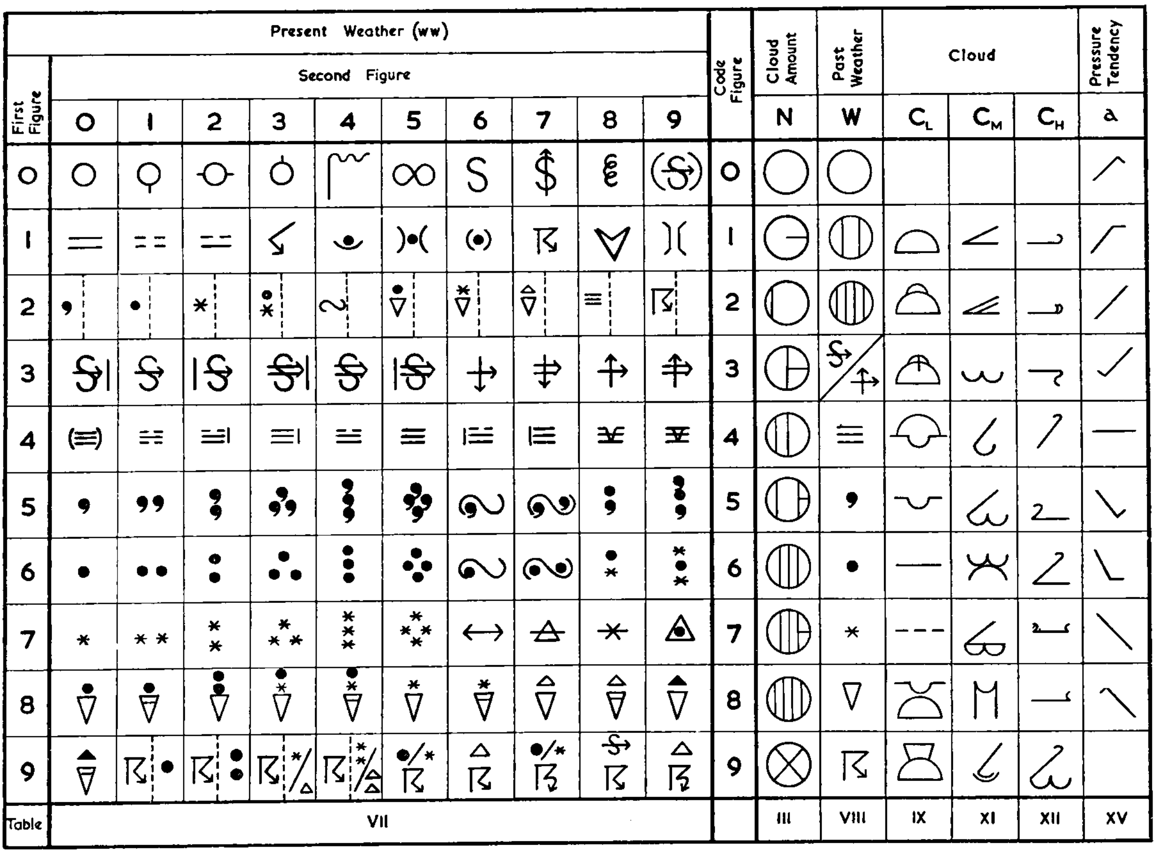


Fig. 12.5. Symbols used for plotting on synoptic charts at forecast offices

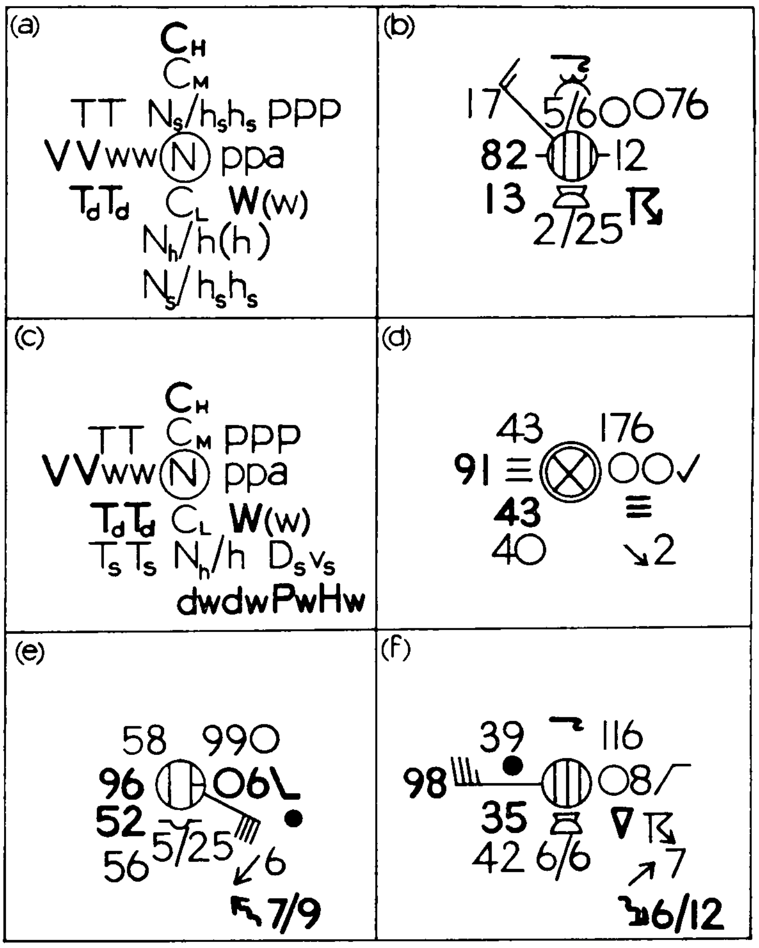


Fig. 12.6. Station model with typical examples
(Bold type represents red ink.)

(*a*) Model for land station report.
(*b*) Land station report ——83314 82029 07617 29563 13212 82925 85460.
(*c*) Model for ship report.

Ship reports:

(*d*) — ——— ——— 90000 91454 17643 9//// 32300 04043.
(*e*) — ——— ——— 51140 96036 99058 55500 56606 85625 05652 11379.
(*f*) — ——— ——— 62735 98918 11639 69603 17108 04235 18162.

When ww = 20–29, 91–94, the symbols to the left of the pecked line represent 'past hour' phenomena, and are plotted immediately to the right of the past weather symbol (W) in black.

Drawing the Isobars

In the absence of fronts, the drawing of isobars on a synoptic chart is like drawing contours on a survey map, lines of equal pressure replacing the lines of equal height. In drawing isobars the following should be borne in mind:

(*a*) Isobars are always simple curved lines with loose ends at the edges of the chart, or simple closed curves.
(*b*) They must never cross, touch or join (except when two ends of the same isobar join to make a closed curve).
(*c*) Everywhere along an isobar the higher pressures must always be on one side and the lower pressures on the other, and the sides must never be interchanged on passing along the isobar.
(*d*) The pressures on consecutive isobars must always differ by the same interval on a particular chart, except at a col where they have the same

value in the direction across the col. The interval in common use in forecasting offices in the U.K. is 4 mb. Some other countries however use 5-mb intervals.

(*e*) It is better to start drawing isobars where observations are more numerous, gradually extending them to areas where observations are sparse.

(*f*) Simple isobars are more probable than complicated ones. Isobars should therefore be kept as smooth as possible, consistent with the observations.

(*g*) Use should be made of reported winds in accordance with the relation given by Buys Ballot's law. An isobar should be drawn so that a reported wind blows slightly across the isobar towards the side of low pressure. Adjacent isobars should be spaced to give a pressure gradient in accordance with the observed wind speed.

One method of spacing isobars is to use a geostrophic wind scale* in reverse. With practice, however, the isobars can be drawn by eye with sufficient accuracy, close together where the winds are strong and wider apart, in proportion, where the winds are weaker.

A uniform distribution of observations is preferable to a great number irregularly grouped. In general, ocean observations are scanty enough to make some reference to charts drawn for the previous synoptic hour (or earlier) necessary when drawing isobars. Care should be taken that a continuous process of change is represented on successive charts.

Adjustment of Isobars to Fronts

The above remarks apply generally in the absence of fronts, i.e. within a particular air mass. The correct drawing of isobars cannot usually be completed until the positions of the various fronts are known. This is because the pressure gradient is discontinuous at a front and therefore each isobar changes direction more or less abruptly. A frontal analysis† is a great help in the correct drawing of isobars where observations are few. A rigid rule of drawing the fronts before the isobars should, however, not be adopted. In practice, it is far better for the two processes to proceed together, fronts and isobars being drawn tentatively and later mutually adjusted to give a satisfactory final picture.

The help given by a good analysis in drawing the isobars needs no specific example, but it may be worth while to see how drawing isobars can help to determine the position of a front. Fig. 12.7 shows two ships' observations which give evidence that a cold front must be somewhere between them. Where exactly should it be placed? After drawing the isobars near each ship by deduction from the wind direction and force, an isobar of one particular value based on the observations of one ship will be found to meet the corresponding isobar based on the other ships' observations at a point. If this is done for several isobars it will be found that the points where corresponding isobars meet will lie on a line; this line must be the position of the front. This simplified construction assumes that there is no appreciable curvature of the isobars, but it can easily be extended if one knows where the isobars are likely to be curved.

In theory it is important to draw isobars as correctly as possible near a front, since this makes for a better estimate of the speed of the front. But in practice

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* *See* page 37.

† The reader's attention is drawn to the remarks in Chapter 13 about plotting the analysis issued in the Atlantic Weather Bulletin for Shipping.

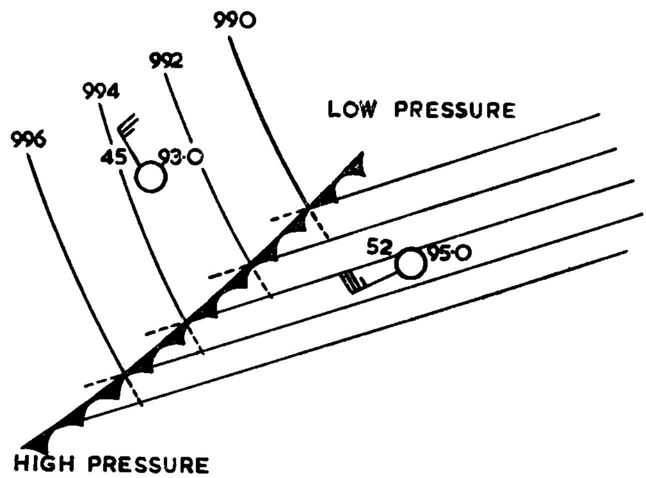

Fig. 12.7. Locating a front by the intersection of isobars

such precision is not often attainable from ship's observations, unless a large number of them is available, and estimates of the speed of a front must to some extent be based upon its recent behaviour.

Use of Pressure Tendencies

The barometric tendency recorded aboard a moving ship cannot be compared directly with the same observation from a land station, until a correction has been applied to it which removes the change of pressure due solely to the ship's own change of position in the last three hours. This can be done quite easily. In Fig. 12.8 the full lines represent isobars on the synoptic chart and **A** is the plotted position of the ship. Knowing the course and speed of the ship, which are given in the synoptic message, lay off from **A** a distance **AB** so that the vector **BA** represents the distance run by the ship in the past three hours. If there were no change in the position of the isobars during the three-hour period the true barometric tendency measured at position **A** would be zero, yet solely on account of its movement the ship registers a tendency expressed by the difference of pressure between **A** and **B**. In Fig. 12.8 this amounts to a rise of 2 mb. When the isobaric pattern is changing, the true tendency is obtained by subtracting this 'spurious' component from the tendency reported by the ship.

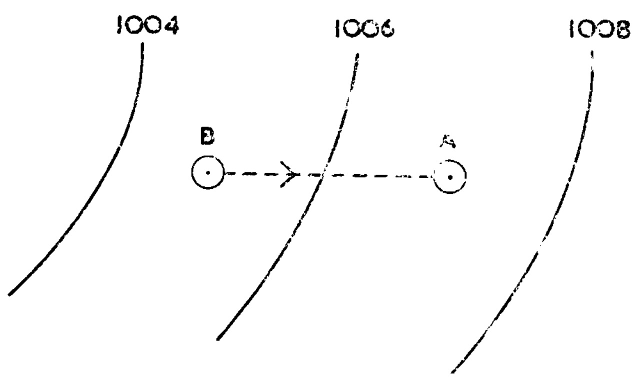

Fig. 12.8. Correction of barometric tendency for course and speed of ship

The corrected tendency readings help to determine the movements of pressure systems, since the barometer usually falls in advance of a depression and rises in advance of an anticyclone. To be more precise, consider the behaviour of the barometer at a position near a depression. Suppose that the pressure at the centre of the depression (its 'depth') remains constant. Then if the centre of the depression moves towards the position, the barometer there will fall. Now consider the case when the depression remains stationary but the barometric pressure at its centre falls; in other words, the depression deepens. We may regard this deepening process as the creation of new isobars at the centre of the depression, the whole system of isobars being pushed slowly outwards. If this outward displacement of isobars reaches the position where our barometer is situated, then the pressure there will fall. Thus, in this case, the falling barometer is not evidence of the movement of the depression but of its deepening. The barometric tendency in any general case thus has two components, one due to the movement or translation of pressure systems, the other due to their change of intensity or development. It is not generally possible to separate these two components, although it would be very desirable to do so.

Where many observations are available, well distributed about the centre of a pressure system, the movement of the latter can be deduced. In practice, when the density of ocean reports is insufficient for this to be done, the movement of pressure systems is estimated by a comparison of successive charts.

The rules given below show that barometric tendency is closely related to pressure changes and developments in depressions and anticyclones, hence it is an advantage to have as many observations of tendency available as possible.

Deepening and Filling of Depressions

(*a*) Frontal depressions deepen after their birth, the rate of deepening usually increasing until some time after occlusion has begun and then decreasing.

(*b*) A depression is deepening if pressure is falling all around the centre, or if the rate of fall on one side is greater than the rate of rise on the other.

(*c*) A depression is filling up if pressure is rising all round the centre, or if the rate of fall on one side is less than the rate of rise on the other.

(*d*) When a new deepening centre moves into the circulation of an old depression, the old depression is either absorbed by the new one or the two centres rotate around each other in a cyclonic sense, i.e. anticlockwise in the northern hemisphere.

(*e*) When a fully occluded depression deepens, it will generally be found that a new influx of either warm or cold air into the system is associated with and responsible for the deepening.

Intensifying and Weakening of Anticyclones

(*a*) An anticyclone is intensifying if pressure is rising all around the centre, or rising on one side more rapidly than it is falling on the other.

(*b*) Conversely, it is weakening if pressure is falling all around the centre, or falling on one side more rapidly than it is rising on the other.

(*c*) The intensity does not change if the pressure tendency is zero at the centre, or if the falling and rising pressures on either side of the centre have the same values.

(*d*) A warm anticyclone in which, by definition, the air at most levels in the troposphere is warmer than the air at corresponding levels outside the anticyclone, is relatively stable and depressions tend to be deflected around it. However it is by no means rare for an apparently stable warm anticyclone to collapse quite rapidly.

(*e*) In a cold anticyclone the air at the surface and in the lower layers of the troposphere is colder than the air at corresponding levels in adjacent regions. Cold anticyclones often build up rapidly in the cold air behind a frontal depression and collapse as rapidly on the approach of another deepening depression.

Identifying the Air Masses

Before the forecaster can use his synoptic chart to make forecasts he must identify the air masses present on his chart. In theory this identification should precede the drawing of fronts, since the behaviour of the fronts depends upon the characteristics of the air masses involved. As was explained in Chapter 9, an air mass is a body of air which has acquired nearly uniform values of temperature and moisture content in the horizontal over an area of several thousand square miles, and through a considerable thickness. This idea is nearer to reality than the reader might at first believe, because there are many source-regions which are suitable for the formation of air masses, such as the subtropical oceans, snow-covered continents, ice-covered polar seas and some of the desert regions. Whenever a supply of air remains over one of these regions for a few days, a distribution of temperature and moisture content in the horizontal, sufficiently uniform to allow this air to be characterised as an air mass, is often attained.

In identifying an air mass the forecaster is really probing into the recent history of the air masses, if not actually tracing them back to their source regions. Thus he is most concerned with those physical properties of the air mass which have undergone little change in their numerical values, while the air mass has moved some distance. These physical properties which change little are known as 'conservative' properties. In practice no meteorological properties can be strictly conservative since processes such as mixing, subsidence, radiation, condensation and precipitation are continually happening within any portion of air, and the last three processes in particular tend to destroy most of the conservative properties of an air mass. Nevertheless this conception is a very useful one to the forecaster in identifying air masses, provided that he also keeps a lookout for representative properties, that is, properties which characterise an extensive region of the atmosphere adjacent to the point of observation. A good example of a representative property is an upper-air temperature (provided it is not measured at a front), though such a temperature is not necessarily conservative.

Properties which are not conservative can be used in the analysis, provided due weight be given to modifying influences. For example, surface temperature on land is of limited value as an air-mass characteristic, because it is seldom representative. At sea, however, where it is of value, the modifying influence is the heat exchange between the air and the sea surface. Fig. 12.9 (14th December 1946) illustrates the use of temperature readings combined with other considerations in making an analysis. The striking feature of this chart is the band of warm southerly winds in the eastern Atlantic, on the flank of the European anticyclone. The lower temperatures in north Scotland, Faeröes and Iceland suggest the presence of a front dividing these two air masses. The locating of the front depends on information from previous charts and the drawing of isobars to fit both pressure and wind. Colder air is cutting into the band of southerlies on its western flank, as shown by observations from the Azores and from ships giving temperatures of 55°, 50° and 47°F. It is evident

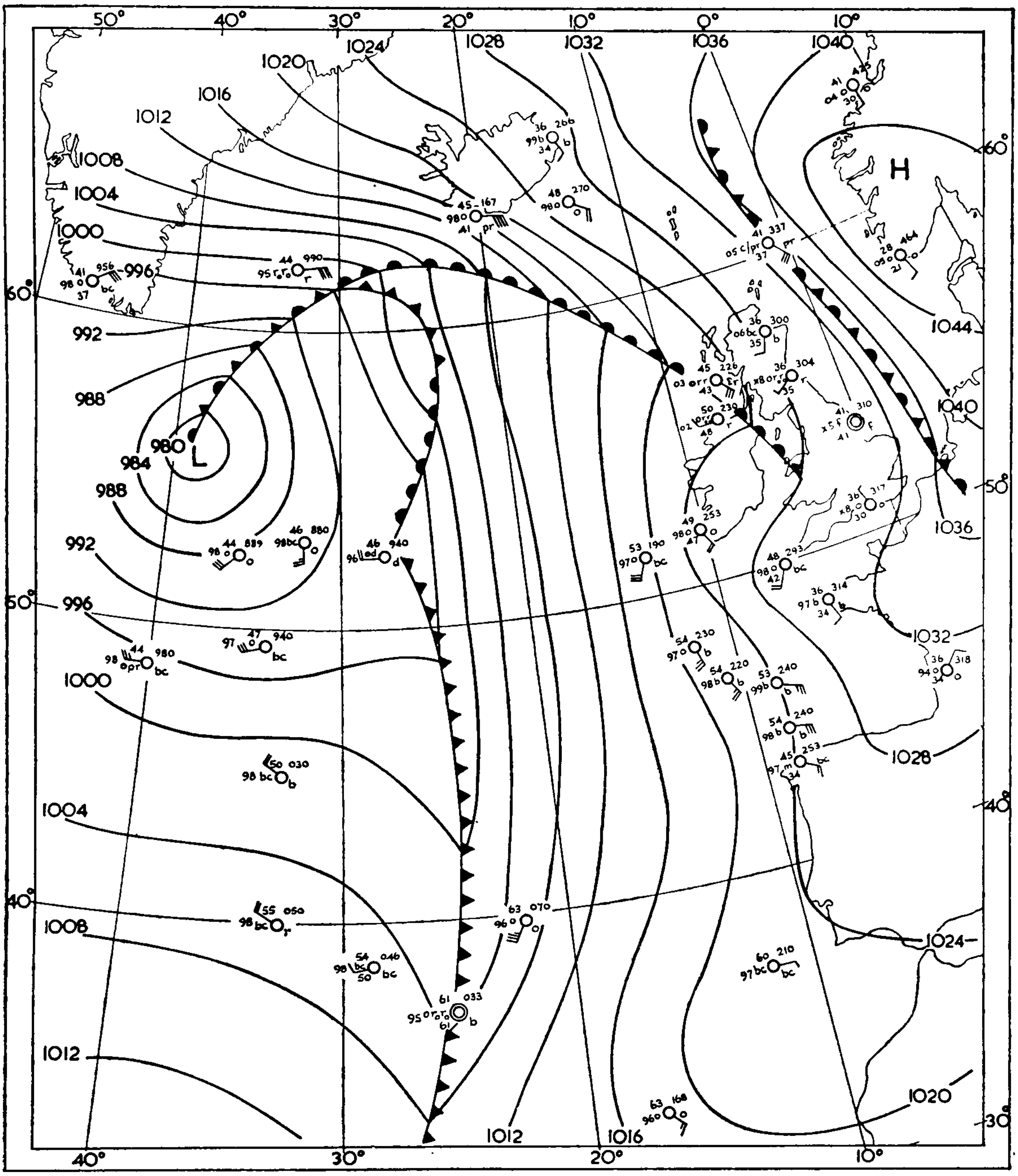

Fig. 12.9. Synoptic chart of North Atlantic, 0600 GMT, 14th December 1946

that these ships are not in the same air mass as the one further east near 40°N, 22°W, which reports a temperature of 63°F. In the south, drawing the isobars to fit the wind fixes the cold front fairly accurately, while the position of the warm front to the north is determined by the run of the isobars, together with the observations to the NW of the British Isles, in Iceland and from the ocean weather ship in the Denmark Strait. The suggestion of a secondary depression at 52°N, 28°W is derived partly from theoretical considerations. The occlusion drawn from a point E of Iceland through the Shetlands to the Low Countries, marks the discontinuity between air derived from an Atlantic air mass and continental-polar air.

Observations of weather phenomena are useful in the analysis, but it is necessary to distinguish between those occurring within the air mass and those in the frontal zone at its boundary. Weather phenomena within an air mass are closely connected with its stability. At sea, where upper air data are not readily available, the stability of an air mass must be judged by its effect on the weather. The word unstable applied to an air mass is a convenient label for a particular class of weather phenomena, comprising showers, thunderstorms, line squalls and turbulence. With a stable air mass we associate stratified low cloud, generally St or Sc, and fog patches; precipitation, if it occurs, is generally in the form of drizzle or intermittent slight rain. In temperate zones, polar or arctic air masses are usually unstable and tropical ones stable. This is due to the modifying effect of sea surface temperature as the air mass travels from its source, to lower latitudes in the case of polar or arctic air and to higher latitudes in the case of tropical air.

Observations of cloud types form a useful indication of the stability of an air mass and indirectly help classify it as polar or tropical. Large Cu and Cb clouds characterise unstable air, whereas widespread St or Sc can only occur in a stable air mass.

Visibility observations are also a help in certain cases. Good visibility may occur in either polar or tropical air masses and one can apply no hard or fast rule. However, if poor visibility is found in association with an air mass identified on one synoptic chart, then this can be of help in identifying the air mass on successive charts.

Fixing the positions of Fronts

Having identified the main air masses, the next problem is to define their boundaries, in other words the fronts, and to relate these to the life history of the system. As was described in Chapter 9, fronts are generally marked by extensive weather phenomena and cloud systems that are dependent upon the type of front. The passage of a front at any place results in a characteristic sequence of changes of most of the meteorological elements. Tables 9.1 (page 112) and 9.2 (page 114) show typical changes at warm and cold fronts in the temperate zones. The changes occurring at occlusions are rather more complex but to some extent resemble those occurring at either a warm or a cold front.

In practice, the forecaster usually has available a sequence of charts with which to determine the changes occurring at each individual station. By making a detailed comparison between the latest chart which he is drawing up, and the charts for a few of the synoptic hours immediately preceding, he can mark the position of the fronts more accurately than is possible by using the latest chart alone and not referring to its predecessors.

Logical Sequence of Charts

A fundamental principle in the analysis of weather charts is that every chart should follow logically from the previous one. The word 'logically' implies that the movement of the fronts should be in accordance with the wind field and that the whole process of change should be a continuous one. Where new features appear on a chart they must be accounted for historically. For example, where a new front appears it should be the result of a front-forming or frontogenetic process already shown on a previous chart. Fronts disappear only as a result of a front-dissolving or frontolytic process.

Importance of Pressure Distribution

A knowledge of the pressure distribution such as is given by a synoptic chart or series of charts at short intervals is essential for the preparation of a forecast, whether for a sea or land station or area, or for a route to be followed by a ship or aircraft.

On the chart can be seen the features described earlier; each LOW is an example of a depression and each HIGH an example of an anticyclone, while we use the terms COL, RIDGE and TROUGH to describe certain other features.

Broadly speaking the main changes of weather in any region of the chart can be explained in terms of the movements of the various isobaric systems. This was recognised in the days when weather charts were first drawn up before the advent of wireless telegraphy. The early forecasters worked largely with isobar patterns until the methods of air mass and frontal analysis introduced by Bjerknes and others were generally adopted. However, a knowledge of the pressure distribution is still necessary, because a forecaster cannot estimate the future positions of fronts, depressions and other features on the map, along with their development, without first knowing as much as possible about the features and behaviour of the pressure systems on the current weather chart.

Requirements of a Good Analysis

An accurate analysis paves the way for a good forecast. An analysis should show:

(*a*) The positions of depressions, anticyclones and other features of the pressure distribution.

(*b*) The extent and identifying properties of the different air masses involved.

(*c*) The relation of the air-mass boundaries, or fronts, to the depressions.

(*d*) Accurate drawing of isobars, especially in the vicinity of fronts.

(*e*) A logical and consistent development from the previous analysis.

Over the greater part of the ocean the number of ships' observations is inadequate for the positions of barometric features to be accurately fixed. In drawing up the chart and in analysing the air mass over a certain part of the ocean, the forecaster may have to place great reliance on some particular ship report. Whereas on land a doubtful observation can usually be confirmed or rejected by comparing it with adjacent observations, this is often impossible over the sea and the observation must be accepted at face value. This is the main reason why a high standard of accuracy is required from ships' observations.

Preparation of Analyses and Forecasts

When the forecaster has completed the analysis of his chart, by drawing up the isobars and fronts as described, he is then in a position to issue forecasts for the region covered by his chart. Bulletins are issued for shipping by national meteorological services in various parts of the world, but as an example here the requirements of a ship in the eastern North Atlantic will be considered. A forecast for shipping in this area is included in the Atlantic Weather Bulletin issued by the Meteorological Office. Part III of this broadcast contains forecasts in plain language for specified areas of the eastern North Atlantic and covers a period of 24 hours from the time of issue. Before he can make a forecast for this area, the forecaster must first make his own mental picture of the pressure distribution anticipated at the end of the 24 hours; this is now done as a routine procedure by drawing up a chart, known as a PROGNOSTIC,* upon which are drawn the forecast positions of the pressure centres, isobars and fronts 24 hours ahead of the most recent main synoptic hour chart. An explanation of the formation of depressions was given in Chapter 9 and some account of the principles used in preparing this prognostic chart are described in Chapter 13. The main problems which confront the forecaster are to decide upon the changes which are likely to take place, each of which is known as a DEVELOPMENT. These changes involve both the intensification and dissipation of existing systems and the appearance of new systems (such as wave depressions or anticyclones, troughs or ridges) during the 24-hour period; and then the forecaster has to estimate the direction and rate of travel of these systems, though these two problems are to some extent inter-related.

Use of the Prognostic Chart

Once the prognostic chart has been drawn, and assuming that events prove it to be correct, the forecasting of the weather for the specified area during the prescribed 24 hours or other period, depends upon a time interpolation of the pressure changes, together with a good knowledge of synoptic climatology. The latter knowledge is needed for deducing such elements as weather, cloud and visibility in relation to the relief (geographical features) of the locality, and to the season, time of day, state of ground and wind strength. Over the sea the problem is to some extent simplified and, as a rule, the temperature of the sea surface (in relation to the air temperature) is the important element to be considered besides the seasonal factors, though the state of the sea surface also needs taking into account, since breaking waves increase the moisture content of the surface layers of the atmosphere to some extent. However, at the present time there are still many reasons which make it difficult to draw a prognostic chart for 24 hours ahead which is accurate in every detail. For instance, ships' observations may be missing from a part of an ocean where (unknown to the forecaster) a frontal wave development is starting. At almost any season in temperate latitudes a frontal wave development may develop into an intense depression well within 24 hours and necessitate the issue of storm or gale warnings over a wide area. This is one of the reasons why synoptic reports from ships can often be so vitally important.

Even in cases where the development is correctly forecast, the true speed and direction of movement of, say, a depression in temperate latitudes may be sufficiently different from the expected values to result in an error of the order of 150 miles in the position of the centre and its associated fronts. In consequence there may be large differences between the wind, weather and cloud actually experienced at a particular locality affected by the system, compared with those which the forecaster was led to infer for the same locality from his prognostic chart.

When a forecast is required for a period of a few hours, changes in the pressure distribution are clearly smaller and less important than in the case considered above, and the forecast problem is more closely related to the factors mentioned earlier in this paragraph. Some problems of local forecasting in which these factors need to be considered by the mariner are given in Chapter 13.

* In the First Edition of *Meteorology for Mariners* the British war-time title of 'PREBARATIC' was used instead of the international title of 'PROGNOSTIC' for a forecast chart.

Use of Upper Air Charts in Forecasting

These paragraphs can be omitted by the reader who is not interested in the meteorology of the upper air, since it does not affect the main arguments in the chapter.

Since the forecaster is dealing in practice with a three-dimensional atmosphere which also partakes in the earth's rotation, it is hardly surprising that the development and movement of surface pressure systems has been found to be dependent upon the pattern of the air flow at higher levels. Within the past 15 years the number of stations making routine measurements of upper air pressures, winds, temperatures and humidities has greatly increased and over Europe, North America and the large part of the North Atlantic covered by the ocean weather stations, the network is now sufficiently dense to enable synoptic charts of the upper air to be drawn. In the United Kingdom and in forecasting offices of many other countries the standard practice is to draw these charts for 0000 and 1200 GMT daily, a separate chart being drawn for some or all of the levels at which pressure has the value of 700, 500, 300, 200 and 100 mb* respectively. On each of these charts the information entered against the stations consists of the actual wind at the level to which the chart refers, together with the 'thermal' wind, the height of the surface above mean sea level and the vertical distance separating the 1000 mb surface from the pressure surface for which the chart is drawn. Temperature and humidity at the chart level are also entered for each station when available.

This information is used to produce charts showing the contours of the actual heights of the various pressure surfaces above the ground (*see* Chapter 9), and also charts which show the contours of equal vertical separation between the 1000 mb and 700 mb, 700 mb and 500 mb, etc., surfaces over the area for which the chart is drawn. The contour lines are drawn at intervals of 60 m through all points where the vertical separation of the surfaces is equal and the lines are interpolated by eye in the same way as when drawing isobars. The charts on which these THICKNESS CONTOURS are shown are known as THICKNESS CHARTS.

A simple deduction from Buys Ballot's law is that the thermal wind direction and strength is related to the gradient of the thickness contours in the same way as the geostrophic wind is related to the barometric gradient. This means that (in the northern hemisphere) the thermal wind blows along a given thickness contour with contours showing smaller thicknesses on its left and that the magnitude of the thermal wind is inversely dependent upon the distance between the thickness contours as measured upon the chart. (In the southern hemisphere this relation also holds, but the thermal wind blows with contours showing smaller thicknesses on its right.)

These charts have been in use for about 20 years and they have been found to be of great value in forecasting. For instance, it has been found possible to amplify the empirical rule that a depression moves in a direction more or less parallel to the isobars in its warm sector. In its new form the rule states that a depression moves nearly parallel, or slightly to the right of, the thermal wind over a region surrounding its centre, when the thermal wind refers to the layer between 1000 mb and 500 mb (approximately from the surface to 18,000 ft over the British Isles). Another way of describing this fact is by saying that a depression is 'steered' by the thermal wind over this central region, and there are

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* For equivalent heights in feet *see* page 11.

theoretical reasons for the correctness of this view. The contours on upper air thickness charts show troughs and ridges, just as on surface pressure charts, although closed isopleths surrounding regions of large thickness (known as 'warm pools') and regions of small thickness (known as 'cold pools', *see* Chapter 9) are less common than on surface charts; in fact the patterns of isopleths on an upper air chart are much simpler than the isobaric patterns at the surface. It has been found that certain regions of upper air troughs on a thickness chart are favourable regions for cyclonic development, while similarly some regions of upper air ridges are favourable for anticyclonic development. On this account these charts can often prove of great value to the forecaster, and are able to supplement and even correct the forecast which he would otherwise issue, based solely upon the series of surface charts available. The reader wishing to study this subject more deeply is advised to read *Weather Map* (Met. O. 595) and *A Course in Elementary Meteorology* (Met. O. 707).

CHAPTER 13

METEOROLOGICAL ORGANIZATION AND THE PRACTICAL USE OF WEATHER BULLETINS BY SEAMEN (including sections on Facsimile and Weather Routeing)

Collection and Distribution of Meteorological Information

This chapter discusses the meteorological organization for forecasting purposes and goes on to give hints to ships' officers upon the drawing of weather maps and the interpretation and use of weather bulletins.

Successful forecasting of the weather depends on a vast organization functioning not only in each country, but also across national boundaries. The weather knows no frontiers. We have seen how a weather map is drawn from a large number of observations taken at fixed times (synoptic hours). Each country maintains within its territory a network of reporting stations. Reports from these stations are transmitted to a central office by telephone, teleprinter or radio. Each country thus collects its own observations and then rebroadcasts them by teleprinter or radio (but mostly by radio teleprinter) in a collective message for the benefit of other countries. To make the observations understandable by all nationalities they are transmitted in an international code. The observations are also made available to the many subsidiary forecast offices within each country. At sea the collection of weather reports has to be dealt with internationally, for ships of many nationalities may make weather observations in a certain area and the coded results must then be transmitted to designated shore radio stations. Ships' observations from ocean areas are of value to many countries, and it is important that they be made available to all. The problem has been solved by assigning areas, internationally agreed, within which ships are asked to report to individual countries through specified shore stations (*see* Fig. 13.1). Each country receiving them broadcasts the ships' reports in collective messages for the benefit of its neighbours.

World Meteorological Organization

International co-operation is essential for the collection and exchange of observations and the issue of meteorological information, not only for shipping and aviation but for all other purposes. This is fostered by the World Meteorological Organization (WMO), which is responsible for establishing international standards and procedures and for preparing the codes and specifications which are the international language of the meteorologist. The World Meteorological Organization is subdivided into Regional Associations which study the problems of particular areas, e.g. Europe, and into Technical Commissions which are concerned with particular aspects of meteorology, e.g. the Commission for Maritime Meteorology.

In order to provide a network of meteorological observations in all oceans, under arrangements made by the World Meteorological Organization, selected ships of most nations voluntarily make observations at routine hours and send the coded results to appropriate centres by radio. Details of the selected ship scheme and practical instruction about making the observations and coding the results for transmission by radio are given in Met. O. 522, *Marine Observer's Handbook*, and Met. O. 509 (Sixth Edition), *Ships' Code and Decode Book*.

The issue of gale and ice warnings, forecasts and weather bulletins by radio is a recognized service to shipping provided by most countries with a seaboard. Before the last war this service was, with a few exceptions, restricted to the coastal waters of the country concerned. Now, however, in accordance with arrangements made by the World Meteorological Organization, the scope of weather bulletins has been increased both as regards content and the area covered. For example, the Atlantic Weather Bulletin for Shipping issued by the British authorities covers the area 35°N to 65°N, 15°W to 40°W and coastal sea areas Biscay, Finisterre, Denmark Strait and North Iceland; it contains storm warnings, a general inference, a forecast, a selection of ship and shore station reports and enough further information for a synoptic chart to be drawn. International agreement is necessary to prevent wastage of effort and unnecessary overlap when weather bulletins are issued for ocean areas, although for various reasons some overlap is almost certain to occur. A map showing the areas within which certain countries are internationally responsible for the issue of weather services for shipping would be very similar to Fig. 13.1. This similarity is not accidental, for the general principle is that a country which accepts ships' reports from any area is thereby responsible for issuing weather bulletins to shipping in that area.

International Codes

For economic and other practical reasons, codes are essential for the transmission and international exchange of the numerous observations which are needed for synoptic meteorology. If one takes the trouble to write down an ordinary weather message in plain language, it becomes obvious that the code is an extremely useful 'shorthand'.

International meteorology, before the last war, had not agreed on one universal code for use on all occasions and in all areas. This step, highly desirable from the seaman's point of view, was taken with the introduction of the Washington Code on 1st January 1949.

Uses of Codes in Weather Bulletins

Seamen need to use codes for the preparation of weather reports and for the interpretation of weather bulletins. The example at the end of this chapter from the Atlantic Weather Bulletin, which is similar to the bulletins issued for shipping in other parts of the world, will be used to illustrate their use. Details of this Bulletin and of the codes and tables referred to here are given in Met. O. 509 (Sixth Edition), *Ships' Code and Decode Book* (Her Majesty's Stationery Office).

Parts I, II and III of the Atlantic Weather Bulletin are in plain language and contain storm warnings, a synopsis of present weather conditions and forecasts for the various areas of the eastern North Atlantic. Part IV contains the analysis in the International Analysis Code (IAC Fleet). Parts V and VI contain a selection of ships' reports and land station reports in code. Parts I, II and III are transmitted first, since these contain the most important information which is essential to all ships. Parts V and VI are transmitted next, as from these the rudiments of a weather map can be drawn aboard the ship. Some time later Part IV is broadcast, giving more detail for completing the weather map.

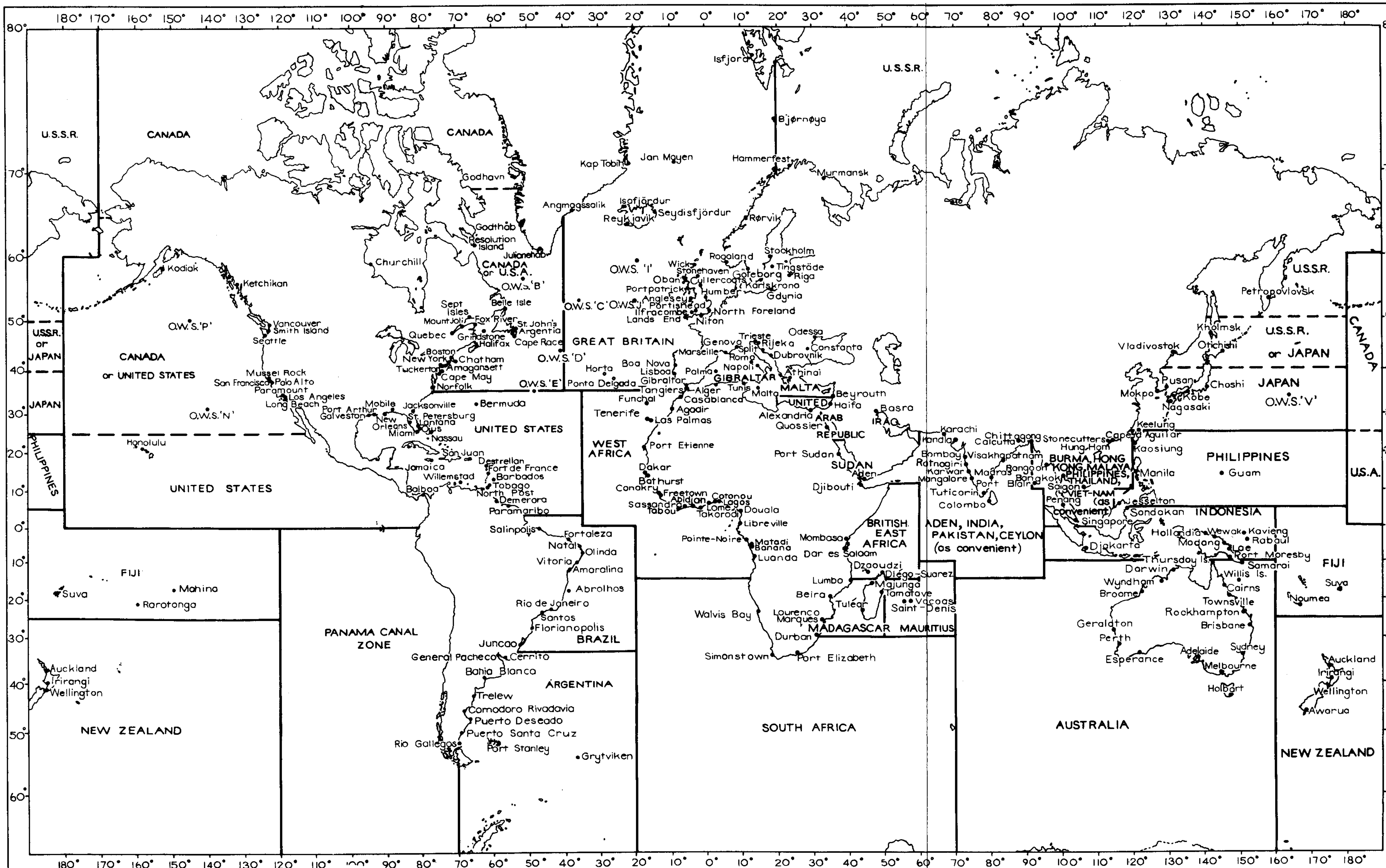

Fig. 13.1. Areas in which observing ships should transmit their radio weather messages to the appropriate stations

(The delineation and naming of the areas and the positions of the radio stations shown on this map are intended for illustration and are not necessarily up to date, for they are liable to change from time to time. Information about the areas and radio stations is regularly revised in *Admiralty List of Radio Signals*, Volume III)

Part V contains a selection of ships' reports in the area, in the form of the first five groups of the SHIP code (FM21C)*:

$YQL_aL_aL_a \quad L_oL_oL_oGG \quad Nddff \quad VVwwW \quad PPPTT.$

Part VI contains land station reports in the form of the first four groups of the SYNOP code (FM11C):

IIiii Nddff VVwwW PPPTT.

Plotting Ship and Station Reports on the Synoptic Chart

As Parts V and VI of the Atlantic Weather Bulletin are received before Part IV, it is a good plan to plot these messages on the chart as soon as received. Station index numbers are usually printed on synoptic charts, but where this is not the case they can be found in the *Admiralty List of Radio Signals*, Vol. IV.

The working chart for use with the Atlantic Weather Bulletin for Shipping is Met. form 1258, copies of which are supplied free to all British selected ships. They may be purchased by applying to the Secretary, Meteorological Office (Met. O. 10a), ~~Eastern Road~~, Bracknell, Berkshire. Working charts for use with Bulletins issued by Meteorological Services in certain other areas may also be obtained from Port Meteorological Officers established at the larger ports of the United Kingdom (*see Admiralty List of Radio Signals*, Vol. III) and various other countries.

Alternatively the mariner can prepare his own working chart on tracing paper, using a map or small-scale navigational chart of the area concerned. Measurements of geostrophic winds cannot usually be made direct from a working chart prepared in this way, since the projection and scale will both differ, as a rule, from that for which the standard geostrophic scales were drawn up. Blank synoptic maps are usually printed on the gnomonic or stereographic projection for polar regions, on the Lambert conformal conic projection for middle latitudes and on Mercator projection for equatorial latitudes. If the working chart is traced from a chart drawn on Mercator projection, no very serious difficulty for the mariner should occur in middle or high latitudes. It would merely result in the shape of the isobars being somewhat distorted. (*See* also page 37 re use of geostrophic wind scale.)

The specimen chart facing page 184 has been plotted and drawn from the observations referred to in the example on pages 183 and 184. By referring from this example to the chart, the method of plotting can be understood. This method is a simplified version of the one used at forecast centres, where more detail is available than given in the Weather Bulletin for Shipping. Those wishing to learn the full method of plotting should consult Met. O. 515, *Instructions for the Preparation of Weather Maps*, published by Her Majesty's Stationery Office.

The position of the land or ship station is marked by a small circle known as the 'station circle'. The various elements contained in the message are then plotted in the relative positions shown in the station model (*see* Fig. 13.2).

The wind direction is shown by an arrow flying with the wind and speed by the number of feathers or pennants on the arrow, each full feather representing 10 knots and each half feather representing 5 knots. A pennant represents 50

* Note: Details of all meteorological codes are liable to change from time to time to meet the changing techniques of the meteorologist. For up-to-date information *see* Met. O. 509, *Ships' Code and Decode Book*, and *Admiralty List of Radio Signals*, Vol. III.

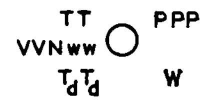

Fig. 13.2. Station model used for plotting the Atlantic Weather Bulletin

PPP=Barometric pressure (in mb omitting initial 9 or 10).
TT=Air temperature (°F).
VV=Visibility (code figures).
N=Cloud amount (Beaufort letters).
ww=Present weather (Beaufort letters).
W=Past weather (Beaufort letters).
TdTd=Dew-point (°F)—available from land stations only.

knots. A north (360°) wind of 35 knots (i.e. corresponding to Beaufort force 8) is indicated as shown in Fig. 13.3.

Fig. 13.3. Representation of a north wind, Beaufort force 8

Beaufort letters are used for plotting cloud amount (N), present weather (ww) and past weather (W). Details of the Beaufort notation are given in Table XXX and the method of using this notation for plotting is illustrated in Table XXXI. *The reader is asked to note that all references to tables in this chapter are to the Tables contained in Met. O. 509 (Sixth Edition), Ships' Code and Decode Book.*

Plotting the Analysis

Having plotted the available station reports and ship reports, the next job is to plot the analysis, which will assist in completing the synoptic map. The analysis is based on much more detailed information than is available to mariners, and it gives in effect a summary of the general synoptic situation in the area concerned. Let us take, for example, Part IV of the Atlantic Weather Bulletin given on page 184. This is coded in the *International Analysis Code for Shipping (see Ships' Code and Decode Book)*. All bulletins for shipping use this code if they include an analysis.

The group 10001 indicates 'Analysis follows'. In the second group 33300 the figures 00 indicate that all positions in the message are in the form $L_aL_aL_oL_ok$ and the latitudes are for the northern hemisphere. The first figure 0 in the third group 01406 is an indicator, while the figures 14 signify the 14th of the month and the 06 following shows that the analysis is based on observations made at 0600 GMT. The next group, 99900, indicates that pressure systems follow; in this section a group beginning with an indicator figure 8 is followed by a position group and sometimes a movement group. Taking the first three groups of this section—81373 37496 90940—the figure 8 indicates that these groups contain details of pressure systems; the figures 1 (from the specification table for P_t) and 3 (from the specification table for P_c) denote a 'low pressure system which is deepening'; the pressure at the centre of this 'low' is 973 mb (the last two figures of the group being 73). The group 37496 gives the position as $37\frac{1}{2}$°N 49°W. The group 90940 is a movement group where the first 9 is a movement indicator figure, the 09 gives the direction in tens of degrees towards which the system is moving (i.e., moving east) and the 40 gives its speed in knots.

The group 99911 indicates the beginning of the frontal system section, where each system is preceded by a group beginning with 66, of which the other three figures give details of the front. In the first case, 66457, the 4 (F_t) and 5 (F_i) indicate 'a cold front showing a moderate intensity with little or no change'. The 7 (F_c) denotes that 'the frontal surface has a tendency to form waves'; in practice, this last figure of the group is usually 0 (frontal characteristic unspecified). The following groups 30515, 34497, 37496 are position groups giving the points 30°N 51°W, 34°N 49½°W and 37½°N 49°W. The position of the front is then obtained by drawing a smooth curve through these points. The same procedure is followed for other series of groups beginning with 66.

Next comes the isobar section with indicator group 99922. Each isobar is introduced by a group beginning with 44; the three other figures in this group are the value of the isobar (omitting the thousands figure). A series of position groups follows giving points through which the isobar is to be drawn as a smooth curve.

The message ends with the indicator group 19191.

Sometimes other information may be given between the isobars section and the 'message ends' group. A group 99955 indicates 'tropical system follows'; 99944 indicates significant weather data follows; 88800 indicates wave and/or sea temperature isopleths follow; 77744 and 44777 are indicator groups before and after a plain language section. Further details of these sections are to be found in Met. O. 509, *Ships' Code and Decode Book.*

Notes on the Completion of the Chart

When the analysis message (Part IV) has been plotted the result should be studied in conjunction with the observations from ships and shore stations already plotted on the chart. In some cases there may be discrepancies, owing to the approximate nature of information given in the message or to errors of transmission and plotting. In making allowance for these preference should be given to the observations themselves. The isobars should then be drawn with as smooth curves and as uniform gradients as possible, and also should be made to conform as closely as possible with the pressure readings already plotted. In drawing these isobars, it must be remembered that they often form sharp angles at a front, where experience has shown that the wind usually changes direction suddenly. The apex of this angle is always directed from low pressure to high. Within the warm sector of a low, i.e. between the warm and cold fronts, the isobars are approximately straight, though they are usually curved in a cyclonic sense near the tip of the warm sector.

It will be found best to use ordinary pencil for drawing the isobars. They can then be adjusted readily until a satisfactory result is achieved. Fronts are usually marked on working charts in coloured pencil using ordinary lines. Details are given in Table XXIX.

Information regarding Movement of Pressure Systems and Fronts

The analysis message usually contains groups of the form $md_sd_sf_sf_s$ where m is a Movement Indicator Figure (*see* Table XXXVI), d_sd_s is the direction in tens of degrees toward which the system is moving (fronts are not included) based on its past behaviour and f_sf_s is its speed in knots. This group appears after the position groups. It can be identified only from its form. In addition, movements of pressure systems and fronts are referred to in the General Inference or Synopsis of Weather Conditions (e.g. Part II of the Atlantic Weather Bulletin). In this context information may be based on theoretical considerations as well as previous tracks.

Whether the information regarding movement is given by groups of the form $md_sd_sf_sf_s$ or is included in a General Inference, it will be based either on the past behaviour of the different systems or on theoretical considerations. In practice a combination of these methods is used.

Forecasting the Movement of Pressure Systems by the Method of Extrapolation

This is the method of estimating future movement purely on the basis of past movement.

For example, suppose the centre of a depression has moved from **A** to **B** in six hours, from **B** to **C** in the next six hours and from **C** to **D** in a further six hours. Where will the centre of the depression be six hours after the position **D** has been reached? EXTRAPOLATION suggests that it lies along the curved line **ABCDE** at the position **E** where **DE** is less than **CD**, because from past behaviour the depression is obviously slowing up (Fig. 13.4).

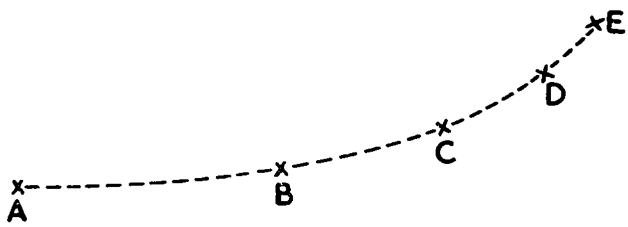

Fig. 13.4. Method of extrapolation for estimating future position of a depression

Extrapolation is an easy and obvious method of forecasting, but it is unreliable unless used with a knowledge of the theoretical limitations involved in any situation. These limitations are taken into account in a number of rules which are useful for practical purposes and are given below.

MOVEMENT OF DEPRESSIONS

(*a*) A depression with a warm sector moves in a direction parallel to the isobars in the warm sector with a speed approximately four-fifths of the geostrophic wind speed derived from these isobars.

(*b*) When the depression has nearly occluded, it moves less rapidly.

(*c*) Occluded depressions tend to move slowly or to become stationary.

(*d*) All depressions move from areas where the pressure is rising to areas where it is falling. The direction of movement as a rule is towards the area of greatest falling barometric tendency. If pressure changes about the centre of a depression are symmetrical, it must remain stationary. This rule can only be used when pressure tendencies are available.

(*e*) Small depressions caught up in the circulation of a larger system have a movement following the main circulation. For example, secondary depressions have a tendency to move cyclonically around the primary depression. When the 'secondary' is comparable in size and depth with the 'primary', both depressions tend to rotate about each other, dumb-bell fashion, in a cyclonic sense.

(*f*) Depressions tend to move around large, warm anticyclones which are well established, in the direction of the air flow around their boundaries.

(*g*) A non-frontal depression tends to move in the same direction as the strongest winds circulating around it, i.e. in the direction of the isobars where they are nearest together.

(*h*) Frontal depressions tend to occur in families, each depression following approximately the path of its predecessor but displaced somewhat towards the equator.

MOVEMENT OF ANTICYCLONES

(*a*) An anticyclone or ridge of high pressure separating successive depressions of a family, moves with the depressions.

(*b*) Compared with the movement of depressions, anticyclones are slow-moving (except as in (*a*)).

(*c*) Anticyclones move from areas of falling pressure to areas of rising pressure. This rule can only be used when pressure tendencies are available.

(*d*) An anticyclone forming in an outbreak of polar air behind the cold front of a frontal depression, moves with the mass of cold air, usually towards low latitudes.

MOVEMENTS OF FRONTS

The movement of a front during a period of time, say six hours, is obtained by measuring the geostrophic wind speed at right-angles to the front for several points along the front and then marking off the resulting movements (according to the 'rules') from each point in a direction perpendicular to the front. The following rules are also useful:

(*a*) The speed of a warm front may usually be taken as about two-thirds of the geostrophic wind speed.

(*b*) The speed of a cold front is usually the same as, or slightly greater than, the geostrophic wind speed.

(*c*) For an almost stationary front the direction of motion is determined by the pressure tendency, the front tending to move towards the side where the pressure is falling.

Extrapolation is also useful in estimating the movement of fronts. For example, if in Fig. 13.5, **AA′**, **BB′**, **CC′**, denote successive positions of the fronts at six-hour intervals, then its position after a further six hours might with some confidence be expected to be **DD′**.

Use of Pressure Tendencies

The ship and shore station reports broadcast in Weather Bulletins for Shipping (e.g. Atlantic Weather Bulletin, Parts V and VI) do not include the group containing 'app' (the barometric tendency). This information is used at forecasting centres ashore but is not considered necessary for use aboard ship in connection with weather bulletins. It is not intended that the mariner should make his own forecast, but merely that he should be enabled to make the best use of the forecast and analyses issued to him, and the Weather Bulletins for Shipping are considered to contain sufficient information for this purpose.

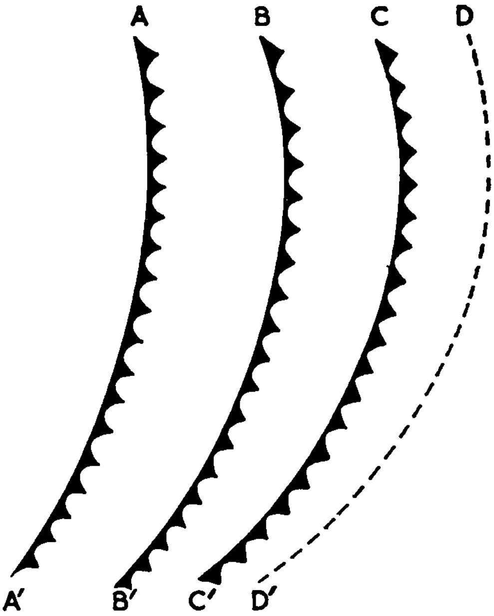

Fig. 13.5. Method of extrapolation for estimating future position of a front

Interpretation of Forecasts

Having plotted his synoptic chart with the aid of the analysis the mariner is then in a position to consider and interpret the bulletin and the forecast much more satisfactorily than if he had no such chart before him.

The forecasts included in weather bulletins for shipping are almost exclusively 'area forecasts'; in many ocean areas, e.g. the eastern North Atlantic, they are given for areas defined by international agreement. The wording of the forecast for each area must therefore be general enough to cover local variations of weather within that area. Forecasts for shipping are usually restricted to weather, wind and visibility, and air temperature in areas where freezing sea spray is likely to be encountered and in some bulletins information about waves is included—all items of practical importance to the mariner.

The mariner will be better able to interpret and make use of a particular forecast when he has acquired a little practical knowledge of the use of synoptic charts for forecasting, such as this book is intended to give him, because he will then have a better understanding than before of the reasons which decided the forecaster to issue it in that form.

Personal Observations and Local Forecasting

Little has so far been said about the value of personal observations of weather aboard ship, which can be used to supplement the official forecasts. The mariner who has just completed a synoptic chart with data from the latest radio broadcast available (e.g. the Atlantic Weather Bulletin), is certainly in a much better position to make use of the official forecasts than his opposite number in another ship who only listens to the forecast. In addition, the mariner who is making observations regularly on passage may be in a position to amend the analysis in the light of his personal observations, when it has gone wrong in certain details. It takes a lot of time to plot a full synoptic chart and to prepare an analysis and forecast, and by the time the radio weather bulletin is received aboard the ship the data upon which the bulletin is based may be about six hours old. For example, Part IV (the analysis) of the Atlantic Bulletin, broadcast to shipping at 1130, is primarily based upon the 0000 synoptic chart. By the time it is received aboard the ship part of the analysis may prove to have been incorrect for some reason, such as a front moving faster than was anticipated, or a ridge of high pressure building up unexpectedly over the area, though caution should always be observed before deciding that an analysis needs amendment. Whether he be afloat or on shore, anyone who tries to relate his own observations to each synoptic chart as it is drawn will soon find himself deriving an increasing amount of information from the chart. This ability to read a synoptic chart will prove of more practical value than any reliance hitherto placed upon ancient rhymes and weather lore, which, in many cases, cannot be justified by any reference to scientific arguments. (*See* also pages 184 and 187 re facsimile maps and weather routeing.)

Since the official forecasts in the broadcast bulletins refer to quite large areas they must be framed in general terms and will seldom contain detail applicable to any small part of these areas. But the mariner with a completed synoptic chart can estimate the probable wind, weather, state of sky and other elements for the next few hours in some detail by relating his ship's position and future movements to the development and movements of the fronts and pressure systems on the chart.

Forecasting of Sea Temperature

Sea surface temperature depends upon several meteorological factors which are to some extent inter-related, and at present there is no method by which it can be accurately forecast at a particular point. A knowledge of the sea-surface temperature is important for the mariner wanting to forecast the development (or dispersal) of fog in his locality.

Away from regions where warm and cold currents are found closely adjacent, sea temperature is usually very conservative and seldom changes as much as 1 °F in a day as a result of meteorological causes alone. A good estimate of sea temperature for a short distance ahead of a ship can often be made from an observation of sea temperature with the bucket, combined with a knowledge of the horizontal gradient of sea temperature in the area, which can be obtained from the appropriate meteorological atlas published by the Marine Division of the Meteorological Office. However, in localities such as the vicinity of the Grand Banks of Newfoundland and off the coasts of Japan, where the horizontal gradient of sea temperature is on the average very steep, an estimate made in this way would have a limited value, since in such an area (as within the Gulf Stream) a ship is always liable to encounter sudden changes of sea temperature wherever detached portions of the warm Gulf Stream and cold Labrador Current happen to be brought close together. Such changes are, of course, largely unpredictable and are not shown in the atlases. The mariner will be aware that these regions also coincide with the areas of maximum fog frequency.

Forecasting of Sea Fog

The forecasting of the time of onset of sea fog is rather more difficult than forecasting radiation fog over land because, with the latter problem, more accurate estimates can usually be made of the quantities involved, such as the rate of fall of air temperature. Also the diurnal range of sea temperature is very small and the temperature of the air overlying the sea only changes markedly if it is moving from a warmer to a colder sea region or vice versa. Formation of sea fog depends on small changes in air temperature, upon the magnitude and sign of the air-minus-sea temperature difference, as well as upon wind speed. It may be slow in forming and in clearing.

Sea fog may be expected when the dew-point of the air is above the sea-surface temperature. From a study of the weather chart, its formation can be anticipated in air which is moving towards a region where the sea temperature is lower than the dew-point of that air; the reason being that on coming into contact with the cold water, that air will have its temperature reduced eventually to its dew-point.

The force of the wind is important for forecasting sea fog. If the wind is strong, usually the fog will be lifted by turbulence into low stratus cloud. Wind of force 3 appears to be most favourable for the occurrence of fog in the Newfoundland region and this probably applies in other regions also. Above force 4 the chance of fog decreases markedly. There is a smaller proportion of fogs with calms and a few cases with forces 6 or 7. On the Newfoundland Banks, however, with a southerly wind, fog can occur with winds of almost any force.

The behaviour of the smoke from the funnel is often a valuable indication. If it hangs about in horizontal streaks, it shows that an inversion of temperature, caused by considerable cooling of the lower layers of air, is already present. This condition may lead to shallow fog formation.

The chance of any marked change of wind direction should be estimated from barometric readings and tendencies, and general weather signs and conditions. If the wind is likely to change so that warm and moist air is replaced by cooler and drier air, the chance of fog will be reduced or eliminated altogether if the change is rapid. In general this means that when the wind starts to blow from a more northerly direction in the northern hemisphere, the chance of fog forming is reduced, or that fog already formed will be quickly dispersed. The effect is particularly noticeable in the vicinity of the Grand Banks of Newfoundland.

The above remarks apply to the normal sea fog of the oceans. In the well-known foggiest regions of the world, fog is frequent throughout the year, in most cases because the sea temperature nearly always is below the air temperature and dew-point. A knowledge of this fact will help the seaman to anticipate fog, even if he has no special experience of the region. In principle, a change to a wind direction which brings air from a warmer oceanic region, at any time of the year, will be favourable to fog formation.

Practical Value of Weather Information to the Mariner

In the open ocean the practical use that the mariner can make of a radio weather bulletin, whether it be associated with an analysis and a plotted synoptic chart or not, depends on circumstances. There is little doubt that with the bulletin and the synoptic chart before him, he should have a reasonably good picture of the existing and impending weather for a period of about 12 hours ahead. For a relatively high-powered ship there would thus be quite a few occasions when a timely alteration of course or speed could be made to avoid bad weather, or to find favourable instead of unfavourable winds. The same may apply, in somewhat fewer cases, to a low-powered vessel. A bold alteration of course in an unfavourable situation may certainly increase the mileage steamed, but by gaining favourable weather fuel may be saved and there may be less risk of damage to the ship and cargo, and less wear and tear on crew and passengers. The same procedure is nowadays frequently used with success by civil aircraft making the crossing of the Atlantic from east to west when strong head winds are anticipated. Fig. 13.6 shows one practical instance where such meteorological navigation proved of definite value to a ship. When the arrival of dirty weather is confidently anticipated, seamanlike precautions can be taken to see that everything is snug aboard the ship in advance of its arrival. (*See* also section on 'weather routeing' on page 187.)

A forecast of foggy weather may justify an increase or easing down of speed so as to fit in with a certain tide, and the owners may be advised by radio accordingly. The action which can be taken must depend very much upon the skill and experience of the navigator himself and the information available.

The practical value of gale warnings and forecasts to all shipping when operating in coastal waters is obvious. Even if there is no bulletin and no official forecast available, the mariner can often do much in the way of anticipating bad weather by a judicious study of his meteorological instruments, and by personal observation of the existing and past wind and weather, in association with 'weather sense'. It is nearly always possible to obtain, by request, radio weather messages from other ships in the vicinity to supplement the information available to a solitary observer aboard a ship.

Example of Atlantic Weather Bulletin for Shipping*

From Bracknell Weather to All Ships, 0600, 14th March 1964.

PART I†. STORM WARNINGS

Storm Warnings. At midnight last night depression of 978 mb, centred 38N 54W was moving east at about 30 knots and is expected to be centred near 40N 40W by midnight tonight. Winds reaching storm force 10 are expected up to 300 miles from this centre in its south and west quadrants.

PART II†. SYNOPSIS OF WEATHER CONDITIONS

At midnight last night a depression of 965 mb centred 52N 21W was moving steadily north. Another depression of 978 mb centred 38N 54W was moving rather quickly east. Centres expected to be about 58N 21W and 40N 40W respectively by midnight tonight with little change of central pressure. Frontal trough moving north-east across Ireland and Biscay.

PART III. FORECASTS

Forecasts from 35N to 65N between 15W and 40W, Denmark Strait, North Iceland, Biscay and Finisterre for next 24 hours (*see* map on page 53 of Met. O. 509, *Ships' Code and Decode Book*).

Denmark Strait, North Iceland.—Wind south-east to east force 5 or 6 freshening force 8 in east of North Iceland later.

West Northern Section.—Wind south to south-west force 5 or 6 locally force 8 in west of section veering slowly west to north-west force 6 or 7 over most of section later. Wintry showers. Visibility mainly moderate but poor in showers.

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* As from 1st August 1964, the words 'wind,' 'force,' 'millibar' and 'visibility' have been omitted from the Atlantic Weather Bulletin. This is to enable the forecast to be given in a clearer and more concise form in the time available.

† Parts I and II are sometimes combined when time will be saved in transmission.

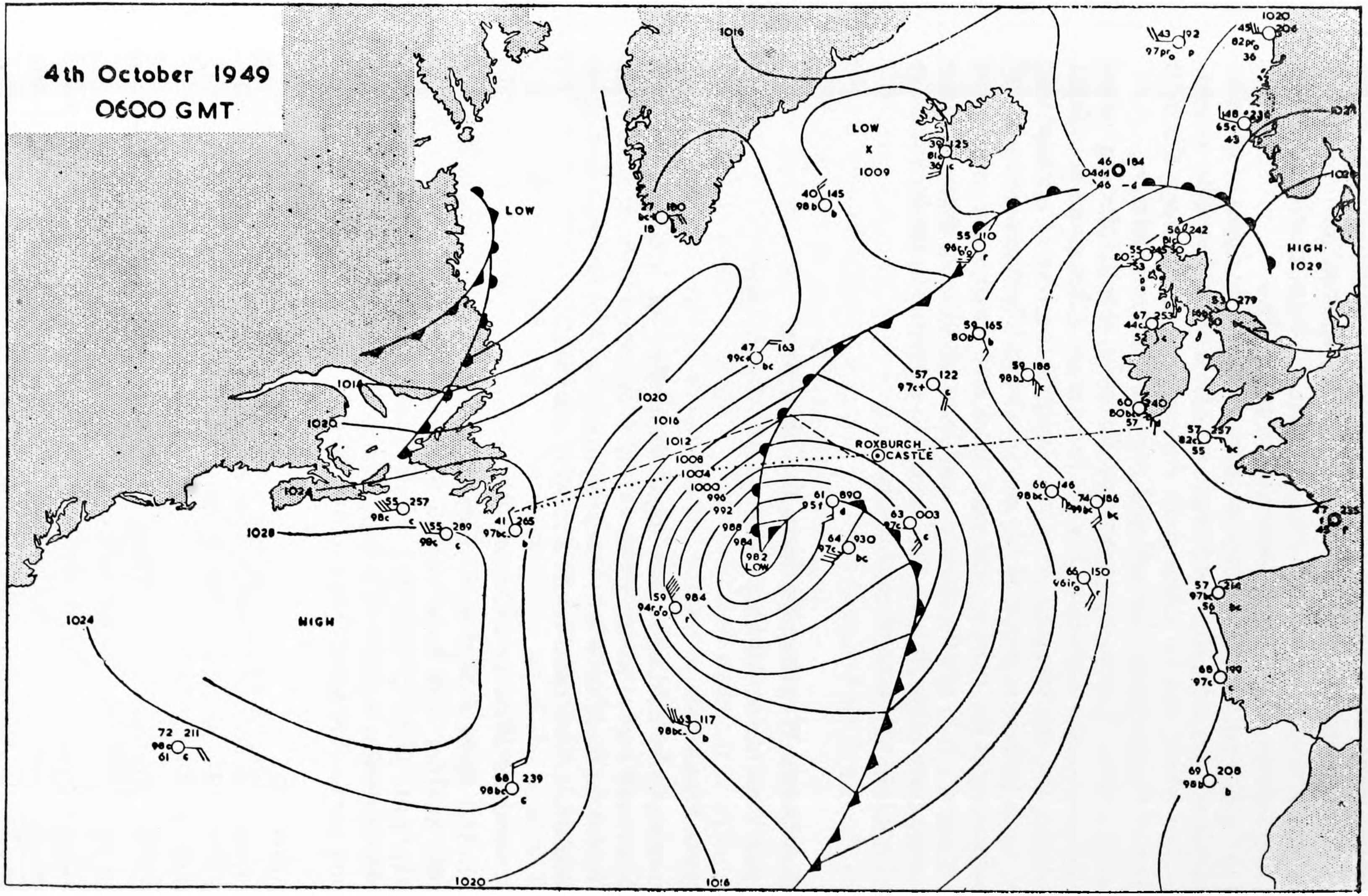


Fig. 13.6. Synoptic chart of North Atlantic, 0600 GMT 4th October 1949

The *Roxburgh Castle* (Captain J. D. B. Fisher) was *en route* from Liverpool to Newfoundland. The normal great circle track was followed as far out as 30°W. On receiving the analysis broadcast in the Atlantic Weather Bulletin, it was seen that there was a small but deep depression moving in a north-easterly direction, so that if the ship had continued on its course the centre of the depression would have passed just ahead of the ship, giving storm force NW winds. The captain therefore changed course to a north-westerly direction until about 150 miles north of the normal route, then followed a great circle track to Cape Race; this secured the advantage of the E and NE winds on the northern side of the depression.

West Central Section.—Wind west to north-west force 7 to gale force 8 in north force 4 in south backing south-east to east and freshening force 6 in south later. Scattered showers. Mainly good visibility.

West Southern Section.—Wind westerly force 5 or 6 soon backing southerly force 4 in east force 6 in west and freshening slowly force 6 in east and gale force 8 or severe gale force 9 in west. Mainly fair at first rain later. Good visibility becoming moderate in rain.

East Southern Section.—Wind westerly force 6 or 7 moderating force 4 or 5 and later backing southerly force 6 in west of section. Scattered showers. Good visibility.

East Central Section.—Wind mainly westerly gale force 8 or severe gale force 9 moderating slowly in south to force 5 or 6. Squally showers. Visibility otherwise good.

East Northern Section.—Wind mainly east to south-east gale force 8 or severe gale force 9 veering south-west to west gale force 8 or severe gale force 9 in south of section. Rain at times. Mainly moderate visibility.

Biscay, Finisterre.—Wind south to south-west force 6 or 7 veering south-west to west force 6 or 7 and moderating slowly later. Rain at first showers later. Visibility moderate becoming mainly good.

Part V. Ships' Reports

($YQL_aL_aL_a$ $L_oL_oL_oGG$ Nddff VVwwW PPPTT—Key Symbols, *see* page 173).

70417	20906	33015	98016	02351
70447	16406	72920	96031	99052
70525	20106	82639	96616	72345
70589	18906	81140	97808	82448
70619	33206	72217	97012	85834
70565	51006	63232	97858	02321
70527	35506	83125	97858	00234
70599	34306	82726	97717	89932
70342	13706	02806	98010	16259
70364	40306	81530	96032	04657
70440	41006	80405	98022	06441
70350	48006	52344	96016	88259
70497	28006	62940	97636	88843
70414	11206	62132	98156	00557
70598	06806	81444	97022	03946
70511	12806	62524	97029	80052
70366	12206	32024	97031	10461
70455	10506	72713	98151	98452
70404	09706	32130	97135	10058
70442	21206	52730	98189	00149
70471	19906	52618	99621	94350
70342	13206	22320	98251	18062
70445	13006	32726	98020	97552
70491	07506	82235	97011	92852
70451	08106	82108	97808	99653

Part VI. Station Reports

(IIiii Nddff VVwwW PPPTT—Key symbols *see* page 173).

03026	11216	70030	03643
03075	81432	56505	12839
03262	81338	24102	08639
03804	82223	61216	93552
03953	81623	48596	81650
03976	81220	62606	82350
01203	00105	85020	30741
01262	11218	89020	33336
06011	81631	96808	11943
04030	11218	65011	95445
04280	83008	70022	85205
07110	82018	58636	99252
07510	71208	70031	07854
08045	82218	60028	03555
08506	23203	82020	15948

PART IV. ANALYSIS AND LATE OBSERVATIONS (For explanation *see* page 174)

Analysis:

10001	33300	01406								
99900	81373	37496	90940	88007	48445					
	81175	62425	10000	88014	42100					
	81266	54216	43620	88085	62316					
99911	66457	30515	34497	37496						
	66150	37496	35446	30405						
	66450	54185	53135	50106	48088	42115	38146	34185		
	66950	56195	54118							
	66450	54118	51065	48047	42068	40085	35125	31165		
	66150	54118	53057	51028	47022					
	66020	64406	68315	70196	73085	75001				
99922	44976	36495	37475	39495	37516	36495				
	44992	32506	35446	39435	42486	40545				
	44008	55537	50485	46555						
	44976	63418	62435	60428	62405	63418				
	44984	66367	65457	61455	59425	61377	64355	66367		
	44968	56218	55245	53216	54186	56218				
44984	50106	53095	54098	58155	60256	59305	55315	51275	49206	50106
44000	42115	48047	51028	57075	62145	66195	69256	69356	71576	69565
	60537	56506	54436	53357	48307	45256	44186	42115		
44008	35395	41367	46407	47386	45325	42266	40196	38146	40085	44015
	48011	55025	60075	65115	70125					
44016	35355	38305	37235	34178	35125	38055	39041	39100		
44016	49080	55020	60035	66075	71035					
44024	55060	60020	65017	70030						
19191										

Facsimile

The value to the shipmaster of an analysis map to help him interpret the written forecast has already been mentioned. Unfortunately the preparation of such a map by hand, aboard the ship—receiving it in morse, decoding it and plotting it—is a time consuming business. Also, in most ships it can only be received during 'single operator' periods.

The facsimile recorder, or radio-telephoto as it is sometimes called, has become widely used by meteorological services nowadays and is by far the most convenient means of disseminating a weather map.

This recorder, connected to a suitable H.F. radio receiver, will by the operation of a switch, reproduce, unattended, in a minute or two, on moist electro-sensitive recording paper an exact copy of the weather map as drawn in the meteorological centre ashore. The recorder can be quite small and can easily be mounted in the average chart room. The advantages of this system are fairly obvious; corrupt groups and transmission errors are eliminated and the map, as received, is drawn by a professional meteorologist. Tests have shown that when the reception of morse or radio teleprinter signals are difficult or even impossible, intelligible weather maps can usually be received by facsimile. Reception is independent of 'single operator' periods of watch.

In addition to the analysis, prognostic (prebaratic) maps showing what the weather situation is expected to look like 24 hours ahead (*see* page 185)—and ice maps and wave maps can be transmitted by this method.

Facsimile maps, not necessarily prepared specially for shipping, but quite intelligible to the average ship's officer, showing surface conditions (an analysis and prognostic) are regularly transmitted by radio by various meteorological services in the northern and southern hemispheres. No doubt more will be provided as the demand increases. The North Atlantic is particularly well served. Analysis and prognostic weather maps covering the eastern North Atlantic are transmitted several times daily, on a regular schedule from the

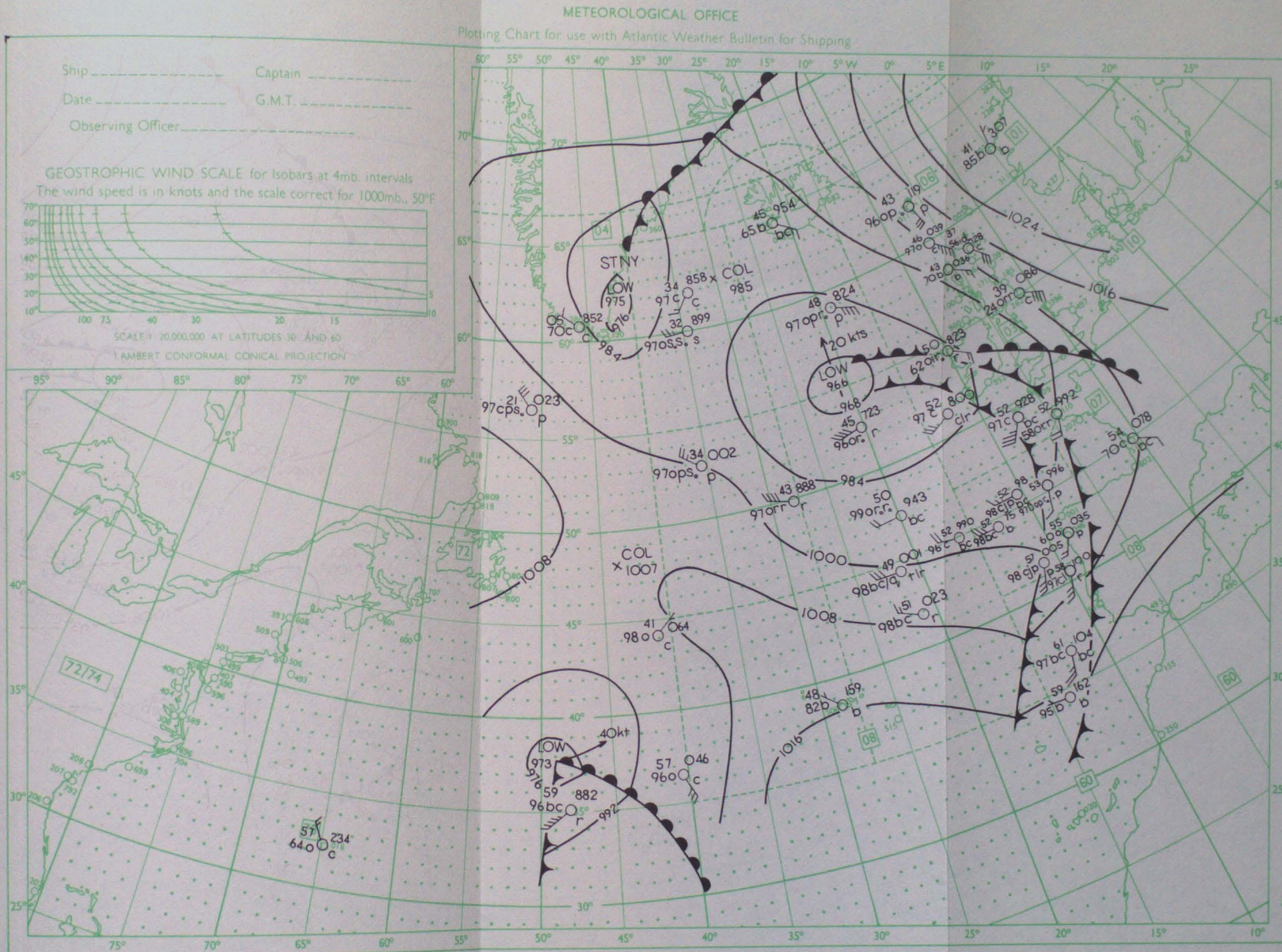


Fig. 13.7. Plotted example of the Atlantic Weather Bulletin for 0600 GMT on 14th March 1964

Note: The isobars drawn on this map are the ones transmitted in the Atlantic Weather Bulletin. To complete the map intermediate isobars (preferably at 4-mb intervals) should also be drawn.

U.K. and from European countries. In the western part of the Ocean, similar maps are broadcast by the U.S.A. and Canada, and they also issue analysis and prognostic maps of wave conditions covering the whole Atlantic, and ice charts from May to September showing the distribution and category of ice in the Gulf of St. Lawrence, Newfoundland, Labrador and Greenland areas.

Similar maps of surface weather conditions in the North Pacific are issued by the U.S.A. and Japanese Authorities.

The facsimile network in the southern hemisphere is not so extensive, but is increasing.

A specimen of a surface prognostic (FSXX) (Fig. 13.8) and of a surface analysis (ASXX) (Fig. 13.9) broadcast by facsimile from Bracknell show conditions in the North Atlantic at 0600 GMT 26th August 1964; these are, of course, very much reduced in size from the original. The prognostic map is of very great value, because it shows exactly, and in many cases better than he can explain in words in a short radio bulletin, what the meteorologist had in mind when he issued his forecast. The intermediate analysis and prognostic maps conveniently fill in the gaps between the times of radio weather bulletins and are thus a ready means of keeping 'up to date' as to the developing weather situation. These two maps were picked at random, and it is interesting to note how closely the analysis, in many respects, fits in with the prognostic, which was issued 24 hours previously. In very difficult situations there won't be quite this similarity!

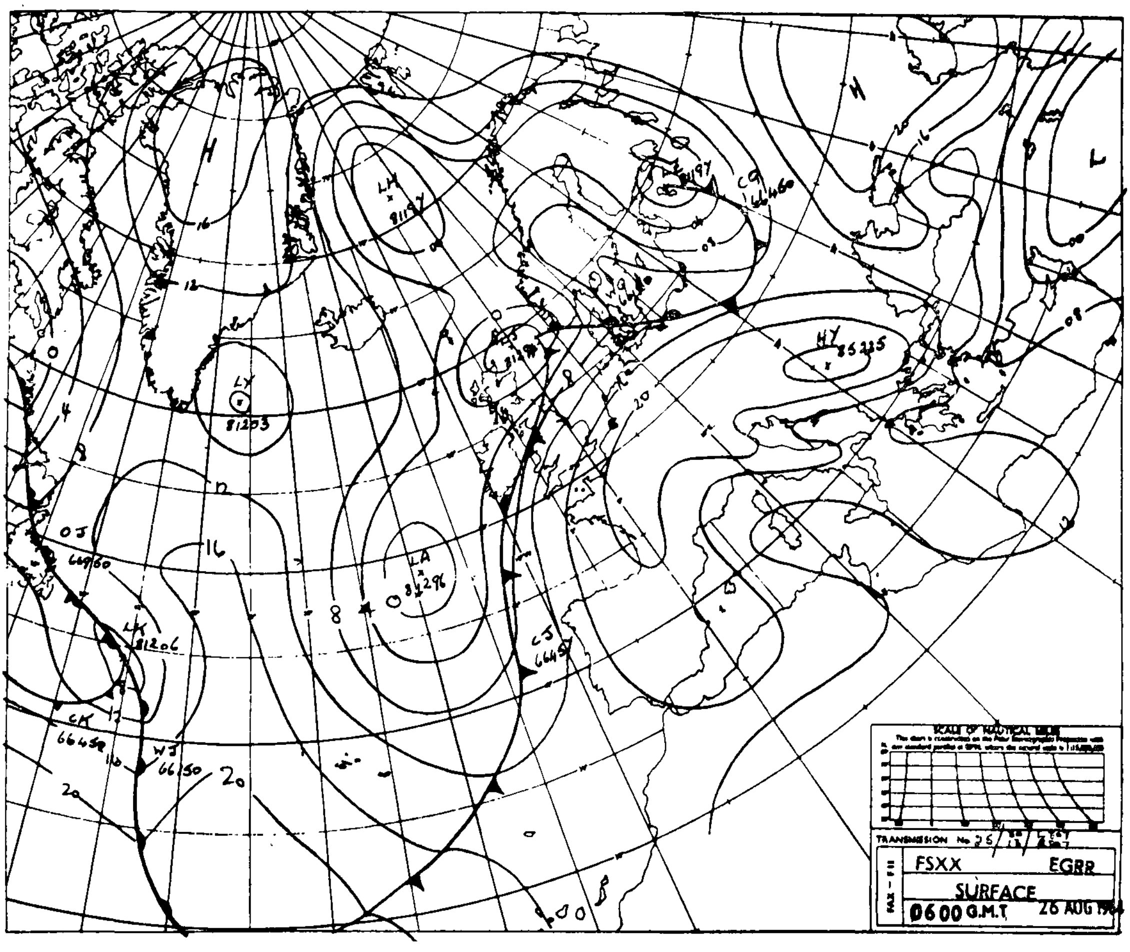


Fig. 13.8. The surface prognostic (FSXX) for 0600 GMT on 26th August 1964 as broadcast by facsimile from Bracknell

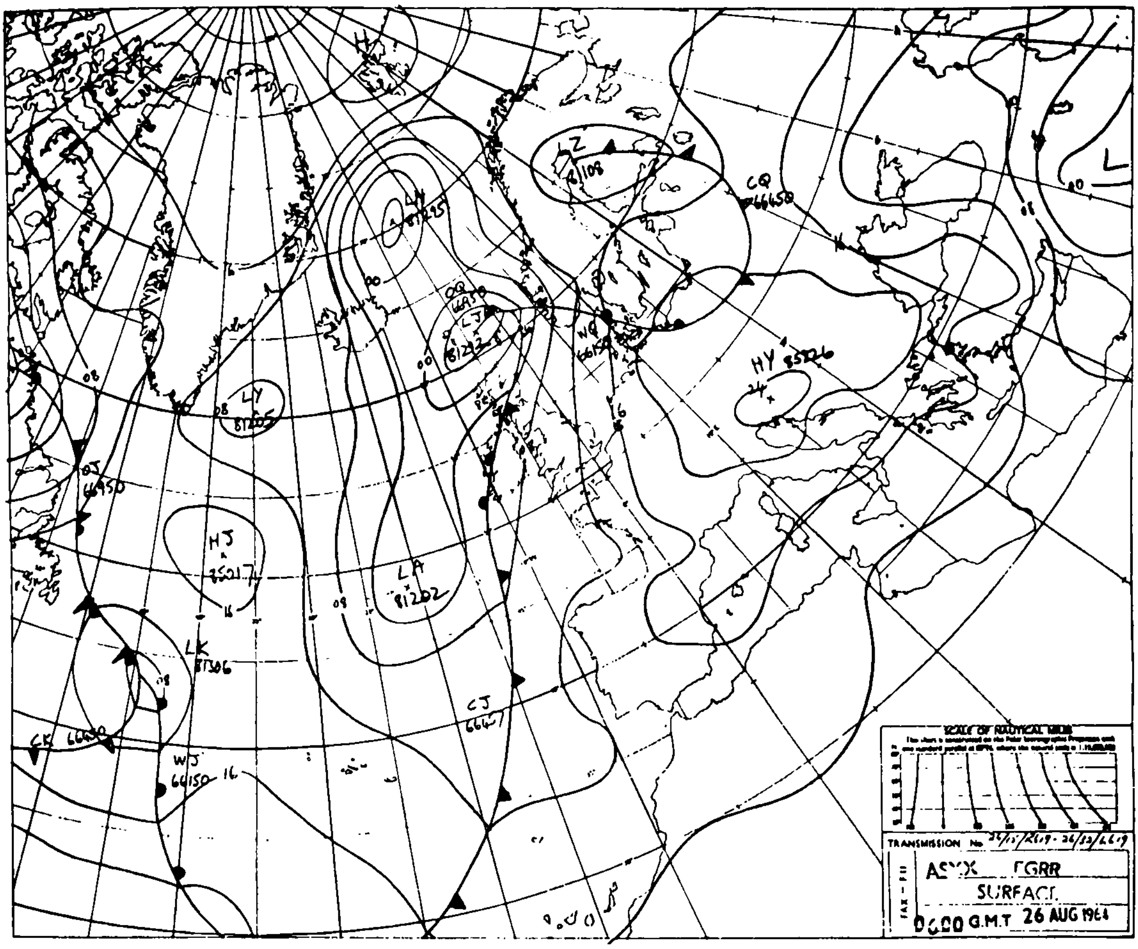


Fig. 13.9. The surface analysis (ASXX) for 0600 GMT on 26th August 1964 as broadcast by facsimile from Bracknell

Even if facsimile comes into general use aboard ship, there will still be a need for the ship's officer to know how to draw a weather map himself, for that is one of the best ways of learning how to interpret the map.

Determination of the Surface Wind Speed from an Analysis or Forecast (Prognostic) Surface Chart

When the isobars on a weather chart are crowded together in any particular area, strong winds, gales or storms are to be expected there; light winds will be found where the isobars are widely spaced. However useful this rather rough and ready approach may be in practice, it is obviously desirable to have more precise information about wind speed when this is possible, and it is intended to show here how this may be obtained when an up-to-date and accurately drawn chart is available.

The first step to be taken in determining the wind speed, is to measure with dividers, the distance in nautical miles between two consecutive isobars, drawn at intervals of 4 mb. (Some increase of accuracy may result by taking the mean distance over three consecutive isobars.) The distance found is now used in conjunction with the table of geostrophic wind speeds on page 187. If the distance is, say, 100 nautical miles and the latitude is 50°, then the geostrophic wind speed is found to be 30 knots. It is important to remember that this is the speed at about 2,000 ft—further steps have to be taken to arrive at the surface wind speed. The speeds shown in the table are correct only when the isobars are straight or very slightly curved. If there is appreciable curvature, the geostrophic wind already found requires to have a correction applied to it—the corrected speeds are shown in the table on page 48. In the present example, with a 30 knots geostrophic wind, if the radius of curvature of the isobars is 300 nautical miles and is cyclonic, then the speed of the wind at 2,000 ft will be reduced to 25 knots. If the curvature is anticyclonic, the speed would be increased to 47 knots. The speed of the surface wind will for practical purpose be $\frac{2}{3}$ of the speed found for 2,000 ft.

Investigations have shown however that when the air is very unstable, the surface wind speed may be as high as $\frac{4}{5}$ of the speed at 2,000 ft. Unstable air conditions may be recognized by the presence of large cumulus or cumulonimbus clouds; these commonly develop in the airstream in the rear of active cold fronts.

The procedure described above may be employed to find the surface wind speed from the charts of other nations provided that the isobars are drawn for intervals of 4 mb.

Table 13.1 Geostrophic wind speeds for 4-mb isobaric intervals

	Distance (nautical miles)												
Latitude	20	25	30	35	40	45	50	60	80	100	150	200	300
0°													
10°				Geostrophic wind speed (knots)						133	89	67	44
20°							136	113	85	68	45	34	23
30°				131	115	102	92	77	57	46	31	23	15
40°			119	102	89	80	72	60	45	36	24	18	12
50°		121	101	86	75	67	60	50	38	30	21	15	10
60°	132	105	88	75	66	59	53	44	33	26	18	13	9
70°	122	98	81	70	61	54	49	41	31	24	16	12	8
80°	118	93	79	67	59	52	47	39	29	24	16	12	8
90°	115	92	77	66	57	51	46	38	29	23	15	11	8

Steep pressure gradients are not experienced in equatorial regions except in tropical storms and accordingly the strong winds shown in the table as being theoretically appropriate to low latitudes are rarely encountered in practice. Geostrophic winds in excess of 130 knots seldom or never occur.

Weather Routeing

Weather routeing can be defined as the art of taking advantage of all available meteorological (and oceanographical) information in order to achieve the most favourable and safest passage for a ship. In its wide sense, there is nothing new in this art. The masters of Arab dhows have used it for centuries in the Indian Ocean; and successful masters of sailing ships all had their own pet ideas about the subject which they kept to themselves. It was not until the 1850's that Maury's *Sailing Directions* provided the mariner with atlases giving somewhat crude seasonal information about the winds and currents of the world, thus helping sailing ships to do much more economical passages than previously. Nowadays, a hundred years of voluntary meteorological work by seamen has provided accurate information about the winds and currents of the world, in atlases and in the *Admiralty Pilots*. The routeing charts now published by the hydrographer contain similar information and also the limits of various loadline zones and most of the conventional steamer tracks on handy monthly maps. Another very useful publication is the Admiralty *Ocean Passages of the World*, giving detailed advice as to the best route to be followed by steamships or sailing ships in all oceans at various seasons of the year; this book also contains climatic and ocean current maps.

Thus, from the climatic or seasonal viewpoint, the mariner has enough information to help him plan the most advantageous passage for his ship. This is recognised in the meteorology part of the syllabus of the examination for Master Mariners.

This seasonal information is probably most valuable in the Indian Ocean and China Sea because of the regularity of the monsoons; in other oceans, also such information can be useful—for example in getting the best out of the Trade Winds or in taking advantage of the Gulf Stream and inshore current respectively off the East Coast of the United States. But, with notable exceptions particularly in temperate latitudes, the day to day weather does not follow a climatic pattern and to get the best out of weather routeing the mariner needs also to know the synoptic situation. Thanks to radio, he now has at his disposal, in all parts of the world, professional meteorologists who are willing to give him the synoptic advice he needs.

In tropical areas, weather routeing to avoid a tropical storm has for many years been the general practice of seamen. Details about this are given in Chapter 11 which also shows that the radio weather bulletins issued by meteorological authorities in areas where such storms are prevalent include advice about the intensity and speed of the storm and thus make it easier for the shipmaster to put these rules into effect.

In temperate latitudes, the frequency and variability of extra tropical depressions throughout the year, the wide area they cover and the speed of their movement makes synoptic weather routeing a much more difficult problem. A tropical storm is relatively small in area and the winds are so violent, it is essential that the ship makes every effort to avoid it for her own safety. In a temperate zone depression, there is not usually much danger if a well-found ocean-going ship does encounter the worst of the wind, because the maximum winds do not normally exceed Beaufort force 9 or 10 anywhere in the storm field. If a wind of force 10 or above is experienced in such a storm it generally does not cover a very large area and a storm warning can usually be issued by radio in time for the mariner to take suitable precaution. The occasions when an ocean-going ship is likely to get into serious trouble in a temperate zone depression are when she is in coastal waters, where there may be the complication of relatively shallow water, lack of sea room and perhaps a lee shore. In the open ocean, it is only when something unusual happens like a bulk cargo shifting or a propeller being lost that the ship gets into real trouble.

In most cases, deliberate weather-routeing of a modern, well-found ship in the temperate zones would be more for economic than for specific safety reasons —i.e. in order to get a more favourable passage, to economise in fuel, to avoid damage to cargo, to minimise the risk of superficial damage on deck or to avoid discomfort to passenger and crew. Ships cost so much to operate and competition is so intense nowadays that the importance of these considerations is obvious. The 'lane routes' in the North Atlantic represent a rather unusual form of weather routeing—the avoidance of ice associated with minimising risk of collision in areas where visibility is often bad.

During recent years, the U.S. Naval Authorities and certain commercial meteorologists in the U.S.A. have instituted a system whereby the masters of some merchant ships, when in the North Atlantic and North Pacific, have been given weather routeing advice throughout the voyage by radio. Similarly the Netherlands Meteorological Service has recently provided regular weather routeing facilities in the North Atlantic to certain Netherlands ships. Both countries claim success with this procedure; exact results are difficult to assess, but the Netherlands claim a three-hour average westbound voyage reduction and some saving in hull and cargo damage in the North Atlantic. In all these cases it is generally accepted that the weather routeing is only in the form of advice to the Master; he has to make the final decision as to the route that he follows, as is right and proper. In both countries the basic principle is broadly the same, but the procedure differs in some respects. In all cases, weather routeing has to be an individual service, provided for a particular ship, at the request of the owners, and with the voluntary co-operation of the Master.

The procedure is broadly as follows. Modern weather routeing technique accepts the fact that waves (both sea waves and swell waves) are the major elements that adversely affect the ship's speed. It is obvious also that waves cause structural damage at sea and, due to the motion of the ship, damage to cargo. The aim of weather routeing is, therefore, to endeavour to advise the Master as to the best route to take on any particular passage, in order to keep the ship out of the worst of the waves and to ensure that she spends the least possible time on the passage and sustains the minimum of damage to ship and cargo. Reduction in weather damage and safety of cargo and crew are as important as spectacular savings of time. It is appreciated that weather routeing a ship is more difficult than routeing aircraft, because of the long initial weather forecast involved. The ship's track needs to be modified at intervals during the voyage, and this entails the maintenance of a careful weather watch on the routed ship so that revisions of the route can be made as early as possible. As the routed ship needs to keep the meteorologists advised as to her position and weather, it is obviously to the advantage of all concerned if she is a 'Selected' Ship.

The aim of the meteorologist is therefore to compute a least-time track for the ship and to do this he has to estimate the distribution of waves and their height and direction for as long as possible into the future along the normal route from the point of departure to the destination. To make the first assessment of this, the U.S. authorities use a five-day long-range forecast; the Netherlands use a three-day forecast.

Prognostic surface weather (and hence wind) charts and prognostic wave charts along the ship's anticipated track are constructed for each of these three to five days. Various formulae and techniques have been successfully tried out for wave forecasting, associated with the use of electronic computers, and there seems to be no reason why these forecasts should not be reasonably accurate on a 24 to 48 hour basis, because wave conditions are so intimately related to the weather systems existing or forecast at the time—whether they are sea waves or swell waves that are being considered. An estimate is then made of the reduction in the ship's speed which is likely to occur due to meeting waves of these specific heights and directions, the anticipated performance of the ship being based upon a graph constructed from information provided from the logbooks of typical merchant ships during many voyages. (Part of this reduction is due not only to the waves themselves, but to the Master's deliberate reduction of speed for the ship's safety.) Ideally, ship performance curves should be prepared for each class of ship and such curves have been prepared for various U.S. Government ships. The curves of a Victory ship which might be considered as representative, in a general sense, of a typical cargo ship are shown in Fig. 13.10. Obviously these performance figures of a ship are governed by her lines, her draught, and her designed speed in smooth water—and the Master of any ship should have a reasonably good idea of her performance in different wave conditions from experience—at least in head seas. After studying these charts and graphs, and the ocean current situation, and information derived from long-range forecasts and climatic charts, the meteorologists draw up for each day of the period under consideration a chart showing isopleths of ship's speed for various courses of the ship through the anticipated waves, taking all the other considerations into account. An example of some of the charts concerned is shown on pages 191 and 192. By continuing this process for each day of the period, it is possible to determine the track that brings the ship closest to her destination at the end of the period—i.e. the least-time track and the Master of the ship is advised accordingly.

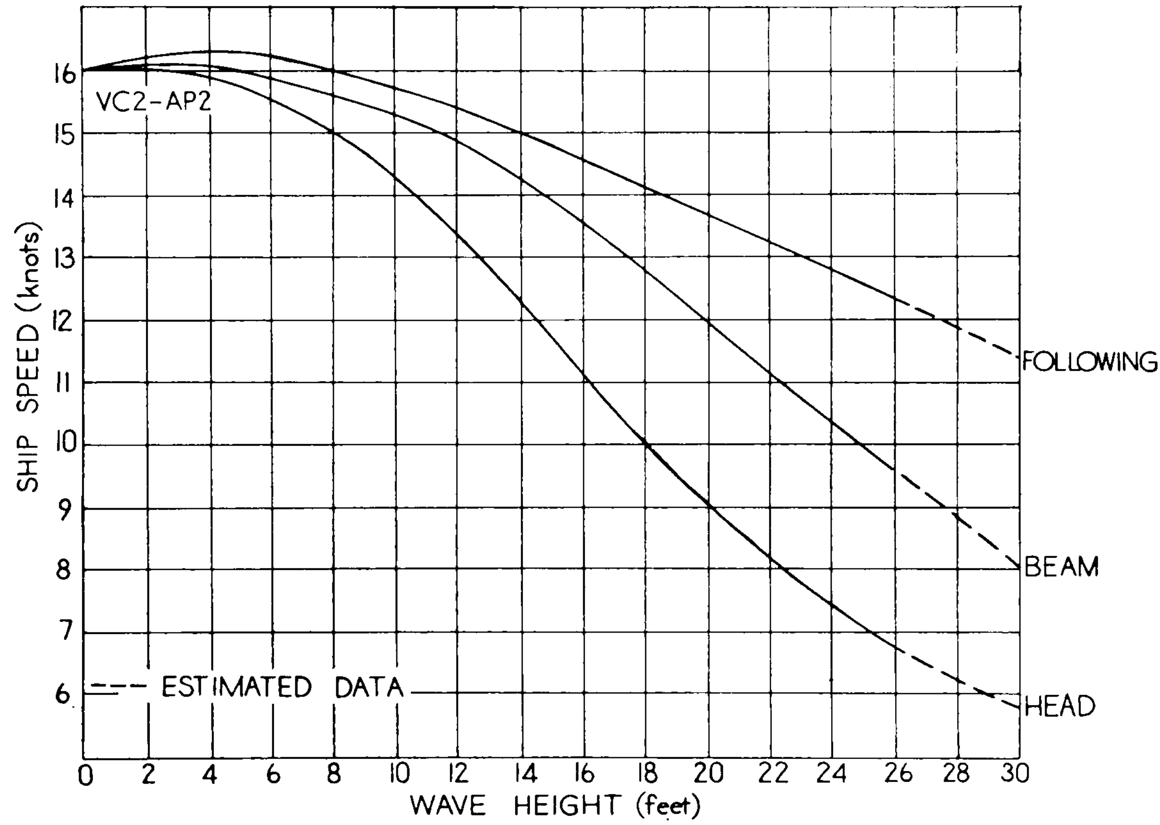


Fig. 13.10. The performance curves of a Victory ship (VC2-AP2)

The Netherlands' system seems to be somewhat less detailed than that described above. Wave maps and prognostic weather maps are prepared and, broadly, waves of 2 metres or less in height are ignored and attention is concentrated on the areas where waves of 3 to 4 metres or more are anticipated. If a major diversion of a ship is anticipated, based on a 3-day forecast, a 'strategic' diversion, less in magnitude, is initiated, based upon a 24-hour forecast and corrected as necessary later.

In both systems at the end of the period, or at any convenient time during the period, the meteorologist will recommend to the Master any amendment to

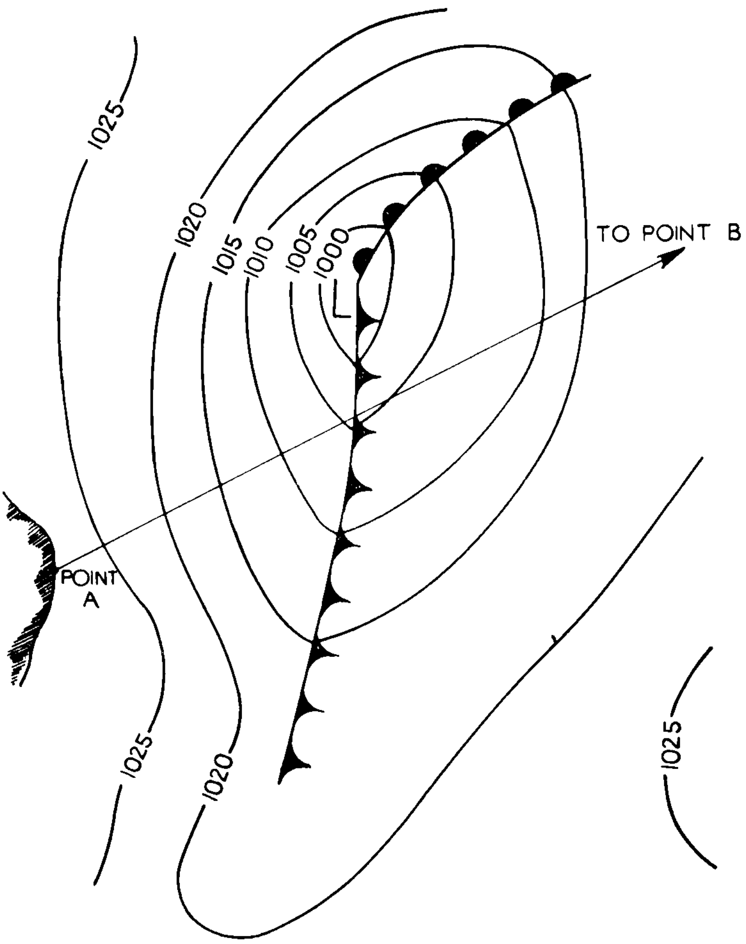


Fig. 13.11. Surface weather chart along track AB

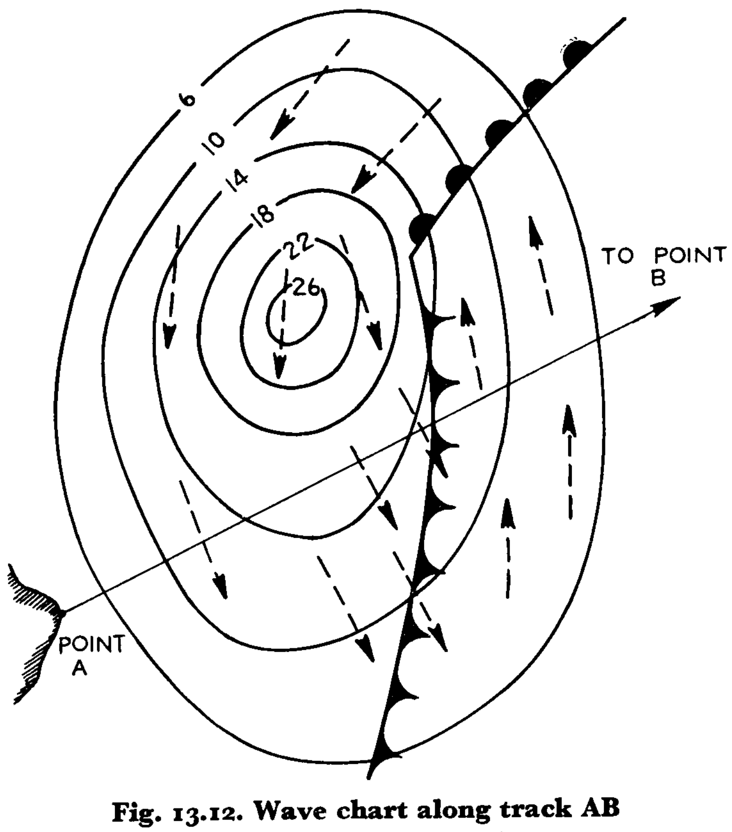


Fig. 13.12. Wave chart along track AB

Wave-height isopleths in feet.

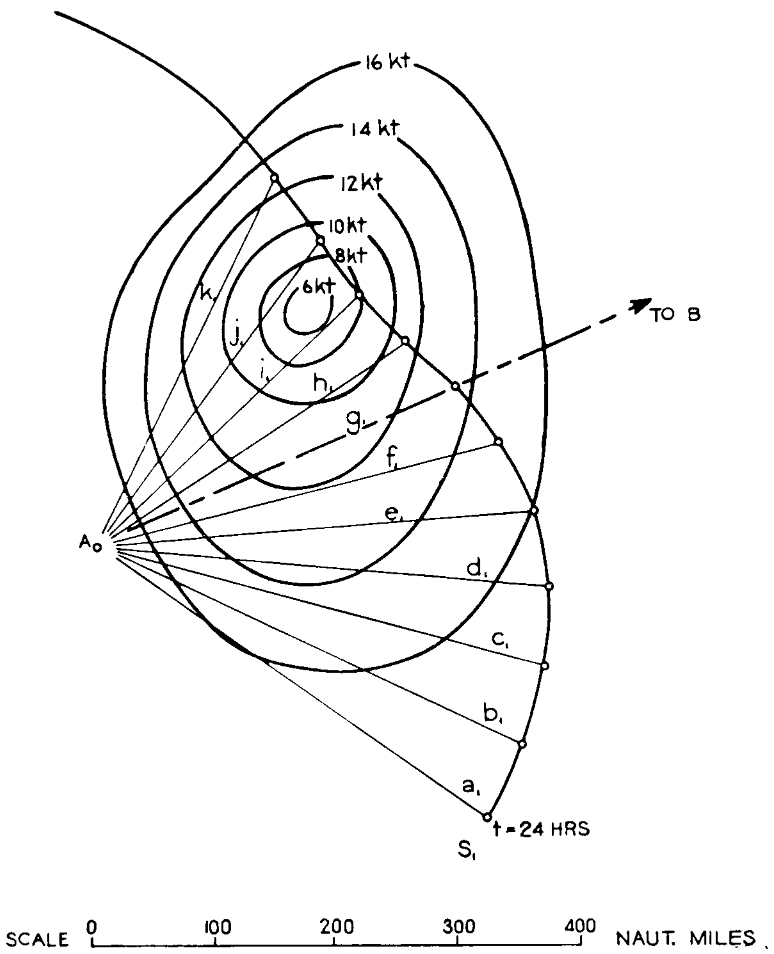


Fig. 13.13. Example of least-time track, first day's travel
Isopleths are ship's speed, straight lines are possible tracks.

the track that he considers desirable; alternatively a reduction or increase in speed may be recommended. As stated earlier, the Master must be free to use his own discretion as to whether he follows the meteorologist's advice. Except in special circumstances, it might be imprudent, in order to avoid strong winds and waves, to route a valuable ship through an area of greater hazard of excessive fog or ice. In all cases, the aim should be the least possible deviation from the shortest route.

In order to get the best advantage out of weather routeing, it seems almost essential that the ship should have a facsimile receiver on board so that the Master can have, readily available, a regular series of analysis and prognostic maps of surface weather conditions and of waves. Armed with these, and with the regular written bulletins that are issued by radio, the Master should be in a good position to interpret the advice given him by the meteorologist and, to confer with him by radio about it as necessary.

It does not seem very likely that the weather routeing described above will be extended to oceans other than the North Atlantic and North Pacific, due to the relative sparseness of the meteorological information available (e.g. from ship reports) in the other oceans.

It seems, however, that the modern shipmaster, if he has kept his meteorology up to date and if his ship has a facsimile receiver so that he can have before him actual and forecast maps of weather and waves, aided by the radio weather bulletins and climatic and ocean current data which are at his disposal, should be able to do his own weather routeing in a simple or modified form fairly effectively. In the North Atlantic, facsimile maps giving analysis and forecast of surface weather are issued every six hours by the U.K. and U.S.A. and similar maps of wave conditions over the whole ocean, every twelve hours, by the U.S.A. Similar facsimile maps are available in the North Pacific. (*See* also pages 184 and 185.) The interpretation of the wave maps in association with the surface weather maps needs practice; the shipmaster is not a trained meteorologist and it is likely that he will need, not infrequently, to consult, by radio, a forecaster of the meteorological service responsible for the issue of weather bulletins for the area in which the ship is navigating. As with all communications, the quality of facsimile reception varies, but with experience the essential features of a poorly defined map can be made use of—here again, radio consultation with the meteorologist may be useful. With these facilities at his disposal, there seems to be no reason why the Master, cannot, on suitable occasions, so navigate his ship as to avoid areas where wave conditions are particularly unfavourable (by altering course or speed or both). As the forecasts in radio weather bulletins for shipping normally cover a 24-hour period, the Master of a ship might, under certain circumstances, ask the meteorologist, by radio, for a special 'extended' (2–3 day) forecast of wind force and direction and of waves along his intended route or he might ask for an estimate as to the reliability of the forecast or seek clarification of some particular aspect of it. In some circumstances he might find it helpful to have a daily consultation by radio with the meteorologist; such consultation only involves the cost of the radio message. Also there may be times when a telephone conversation with a forecast office before sailing will be useful. In other words, the mariner can, if he so wishes, help himself to a large extent to the information which is available to him.

PART V. OCEAN SURFACE CURRENTS

CHAPTER 14

THE OBSERVATION AND CHARTING OF SURFACE CURRENT

Historical

Parts of the general surface circulation were known to the early navigators. Vasco da Gama, who called at Mozambique in 1498 after rounding the Cape of Good Hope, must have experienced the adverse force of the Agulhas and Mozambique Currents, and so might be regarded as their discoverer. Columbus, in his voyages to America, encountered the North and South Equatorial Currents of the Atlantic and stated that he regarded the water movement from east to west as proved. Dampier's *Voyages* published in 1729 includes an account of the currents experienced in his voyages in the latter part of the seventeenth century. He knew of the Guinea Current and gives relative strengths of the Equatorial Current in different parts of the Caribbean. Benjamin Franklin published the first chart of the Gulf Stream in 1770.

By the early nineteenth century seamen had a general working knowledge of the chief trends of current in areas frequented by ships. The first extensive current charts based on ships' observations were those of Rennell, published in 1832, for the Atlantic and part of the Indian Ocean. In 1845, and succeeding years, Maury published his well-known wind and current charts. These early charts and the Admiralty current charts published near the end of the century were based on plotting individual current observations from the logbooks of naval and merchant ships, the average current being shown by arrows drawn by eye estimation. It was not till after 1910 that exact statistical combination of current observations into current roses and vector mean currents was begun.

Observation of Surface Current

For constructing current charts, as many observations as possible are required during many years. The only way of obtaining enough observations is by the co-operation of merchant shipping. The method used in making the observations is to calculate the difference between the 'dead reckoning' position of the vessel, after making due allowance for leeway, and the position by astronomical or land fix. The result is the set and drift experienced by the vessel during the interval since the previous astronomical or land fix.

The current found by this method is that for a mean depth of about half the ship's draught. It will be correct only if the ship's true course and speed through the water are known. Knowledge of the true course involves a precise knowledge of the error of the compass, sound judgment of the leeway made and the leeway allowed for when setting the course, and good steering. An accurate estimate of leeway can only be made by experience, bearing in mind that the greatest effect will be with the wind direction abeam when the ship is light. For the speed through the water a compromise between log distance and distance by engine revolutions, after making due allowance for slip, gives perhaps the best results. Much will depend on local circumstances, e.g. in rough weather, with the propeller often clear of the water, it will not be possible to assess the slip with any degree of accuracy; the same is true to a limited extent when the ship is 'flying light' or when she has been a long time out of dock. It should be possible from time to time to test the accuracy and compute the percentage error of one or more logs by comparison with each other and with runs over a known distance in still water. Judgment is necessary as to the appropriate length of log-line to use at varying draughts. In other words, the dead-reckoning position to use is that which the navigator might expect the ship to be in if all considerations except current were taken into account.

A precise knowledge of the ship's position is essential. The noon position which is so often used depends for its value on the run between observations, and can thus only be considered accurate to a limited extent. When taking a sun sight it is very often possible to obtain a cross with either the moon or the planet Venus. Occasionally Jupiter is visible, whilst a single bearing of a point of land may be used in conjunction with the foregoing and a full land fix should be used if available. Stellar observations at morning and evening twilight undoubtedly give the best positions at sea. In the space of 24 hours a ship may experience two or more distinct currents; observations at short intervals are therefore to be preferred. It is recognized that individual current observations are frequently only approximately accurate, and that in some cases the accuracy may be less than in others. In compiling the current charts, however, such casual errors tend to cancel out in the long run providing that the number of observations in each part of an ocean is large enough. There is another reason why many current observations are required; the day-to-day currents in the oceans are very variable and the general current trend in any small region cannot be determined until there are enough observations to be representative of all the variations, both in direction and rate, which can occur in that region.

The only other way of measuring surface current directly is by using a current meter. When used from a ship at anchor, this measures the rate at which the water passes it in much the same way as the patent log measures the ship's speed. There is mechanism attached for recording the direction of current. Some reference to different types of current meters will be found on page 286.

Use of Current Information

A practical knowledge of ocean currents is necessary to the shipmaster, not only for the safety of his ship, but also to assist in the economical operation thereof. The safety aspect is fairly obvious when one considers the vagaries to which currents are liable, and a study of shipping casualty returns shows how often an unexpected current has contributed to the casualty. Despite the modern aids at his disposal, the prudent mariner must still use lead, log and lookout, and study the information in *Admiralty Pilots* and elsewhere about currents. By making and reporting observations of currents experienced, the seaman not only gains practical knowledge himself, but benefits shipping generally by adding to our statistical knowledge so that we can publish up-to-date information.

An example of the saving of time and money which study of the current chart can effect is given by a voyage made in 1932 from Balboa to Sydney, N.S.W. The shortest route is an uninterrupted great circle, but by steering a rhumb-line to Motu One Island in lat. 15° 48′s, long. 154° 34′w, and thence a great circle to

Sydney, it was found that although the distance had been increased by 170 miles, the Equatorial Current experienced had reduced the percentage slip to 0·6 and given the ship a day's run of, on some days, 40 miles more than her average and never less than 20. The consequent saving of fuel and early arrival of an important cargo well justified taking this route.

In small deep-loaded steamers such knowledge is very necessary. For instance, a ship bound from Colombo to the Straits of Bāb-el-Mandeb in the SW monsoon will, if she has steered a rhumb-line, find when she is to the southward of Socotra that she is steaming against an east-south-easterly set, sometimes as strong as 5 knots or even more. A diversion to the north of Socotra during this season will increase her distance by some 30 miles, but she will be clear of these very strong adverse currents and the heavy confused seas which accompany them.

In June 1923, the steamer *Trevessa* foundered in lat. 28° 45′S, long. 85° 42′E. Two boats were launched and it was decided to sail to the northward and, taking advantage of the south-east trade wind and westerly current, make for Mauritius. This voyage of some 1,600 miles was made in less than a month. At the time of foundering the nearest land was Australia. By proceeding to the southward the boats could have picked up a westerly wind and easterly set which would have also taken them in the direction of safety, but unfortunately this easterly set turns to the northward when still some distance from the Australian coast and subsequently turns west. Thus a knowledge of ocean currents helped in the succour of these distressed mariners.

Charting of Currents

The current charts so far published by the Meteorological Office are constructed for three-monthly periods, except that monthly charts of the China Seas are shown. Many more current observations are needed before monthly charts for all oceans can be constructed. Because of seasonal variations to which currents are liable (*see* page 224), monthly current charts will give a better picture of the currents at any particular time than quarterly ones.

In recent atlases, three types of current chart are given for each quarter (or month) showing: (*a*) resultant or vector mean current, (*b*) current roses and (*c*) predominant current.

Vector Mean Chart (Fig. 14.1)

When an ocean is to be charted for current, it is first divided up into computing areas of 2° of latitude by 4° of longitude. In a few regions where different current trends are known to occur close to one another, the area may be subdivided further. For each area, the vector mean is computed of all observations of currents, the mid-positions of which lie within the area. The vector mean is therefore the resultant rate and direction of current within that area, calculated from the rate and direction of each individual current. Opposite components cancel each other and the vector mean rate is therefore less than the mean rate of all currents irrespective of direction. In areas of very variable current, the vector mean rate is much less than the average rate and may be extremely small. The more constant the current, the closer the vector mean is to the average rate, but the two are never equal, since the current has nowhere a constancy of 100%.

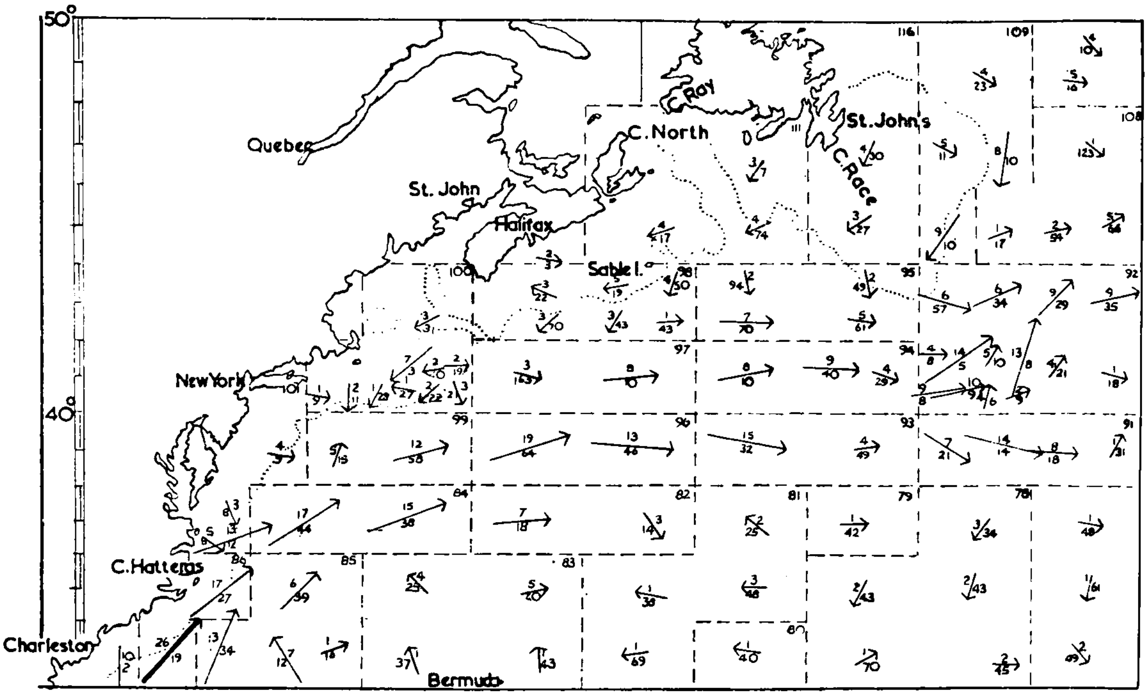


Fig. 14.1. A portion of a vector-mean current chart

The vector mean chart shows the overall movement of water over a considerable period. It defines the geographical limits of various currents and shows the difference in mean direction and rate between individual current trends. It would, for example, be the chart to consult for calculating the average drifts of boats, derelicts, icebergs, etc.

Current Rose Chart (Fig. 14.2)

The rose indicates the degree of variability to which current in a given region is liable. The number of observations is not generally sufficient to provide a current rose for each 2° × 4° area, besides which the charts would have to be

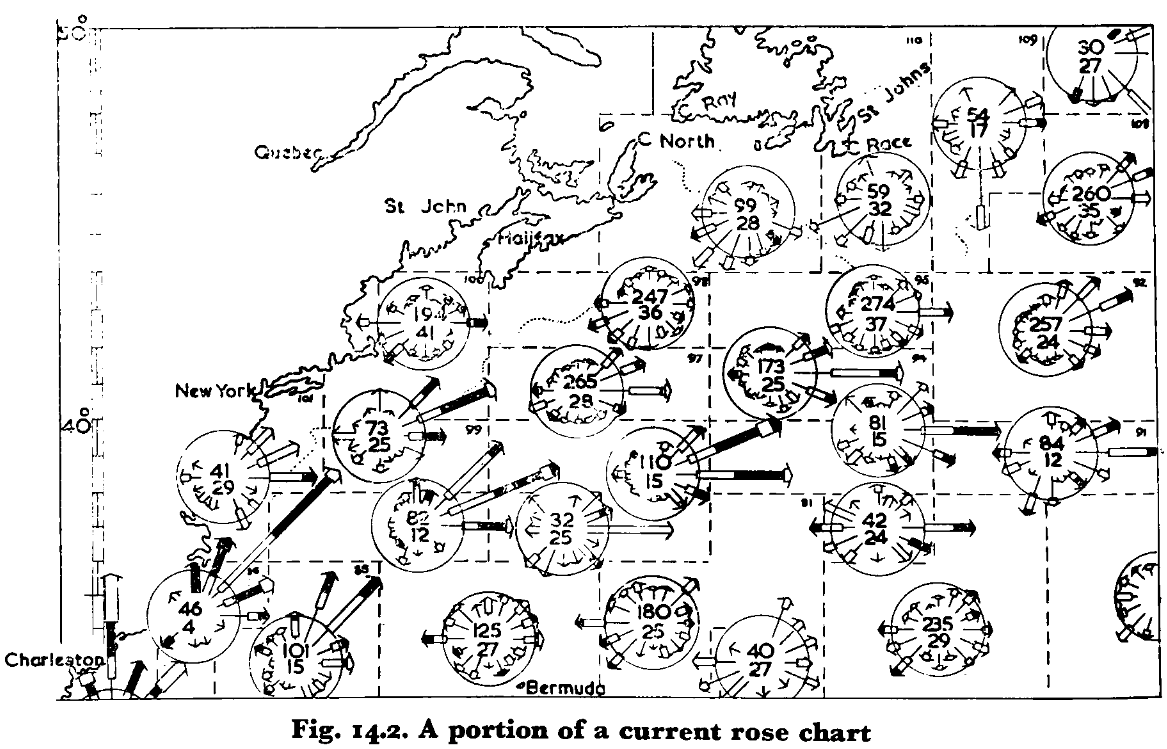


Fig. 14.2. A portion of a current rose chart

on a much larger scale. A number of 2° × 4° areas are therefore combined to form a convenient area, for which one rose is calculated. The rose areas vary in size and shape in different parts of the chart. The principle underlying their selection is that the various currents of the general circulation are separated as far as possible into different rose areas; this separation is not always equally good at all times of the year, as the area covered by some currents varies to some extent seasonally. It has not, however, been found practicable to change the rose areas in the charts for different quarters (or months).

The current rose chart gives all possible information about variability of current. It not only shows the total percentage of current setting in any direction, but also the percentage of currents of various strengths in each direction. Currents of less than one-quarter of a knot are not included in the rose; the total percentage of these, irrespective of direction, is given in the rose circle, beneath the figure showing the total number of observations.

Chart of Predominant Current (Fig. 14.3)

This chart shows the direction of predominant current whenever there is a predominating direction, i.e. where the constancy of current in one direction is 25% or more of all observations. It also gives the average rate of the currents in miles per day actually experienced in the predominant direction. The information given on this chart should be used by the navigator for day-to-day navigation in preference to that given on the vector mean chart, which, as explained above, refers to the resultant flow of water over a long period. *When using the predominant current chart, the rose chart should also be consulted to find what degree of variability from the predominant direction and rate may be expected in the region concerned.*

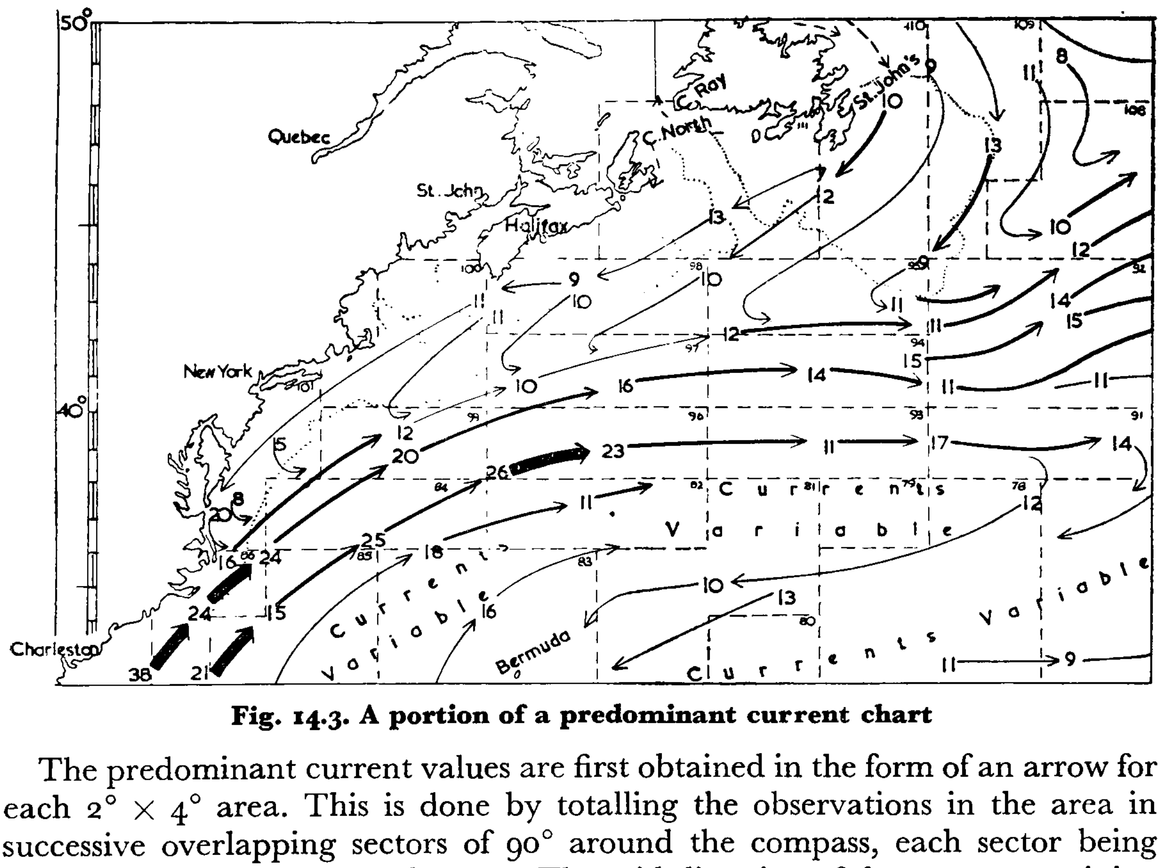


Fig. 14.3. A portion of a predominant current chart

The predominant current values are first obtained in the form of an arrow for each 2° × 4° area. This is done by totalling the observations in the area in successive overlapping sectors of 90° around the compass, each sector being displaced 15°* from its predecessor. The mid-direction of the sector containing

* The displacement of 15° is chosen as being the most convenient one to give the desired result.

the largest number of observations is taken as the direction of the predominant current. The predominant rate is the arithmetical mean of all individual current rates flowing within 45° either side of the selected direction, i.e. the average rate of these currents irrespective of minor differences of direction. The constancy of the predominant current is the percentage ratio of the number of currents in the selected sector, to the total number all round the compass.

On a preliminary chart the predominant currents are plotted as arrows, the constancy of each being appended. The predominant chart is then drawn from this, in the form of a 'flowline' chart, and some smoothing of direction, rate and constancy is made at discretion. Smoothing is necessary where very variable currents occur or where there are few observations.

The flowlines on the chart are of three thicknesses, representing constancies of 25–49%, 50–74% and 75% and over. Where the constancy exceeds 50%, the direction shown on the chart is the most likely one to be experienced at any particular time, this being, of course, more likely where the constancy reaches 75% or more. As the constancy nowhere reaches 100%, no current can be predicted with absolute certainty; this is the reason for the advice given above that the predominant and rose charts should be studied together.

In the regions where the constancy lies between 25% and 49%, more currents flow in or near the predominant direction than in any other *single* direction, so that, in the long run, the predominant direction does indicate the approximate trend of the general circulation. In such regions, on any particular occasion, the chances are against experiencing a current in the predominant direction.

The flowlines of the predominant chart give an approximately accurate picture of the general circulation of an ocean. This does not differ essentially from the circulation shown by the vector mean chart but there may be differences of detail, because the vector mean chart takes account of every observation, while in drawing the predominant chart only those of the 'most frequented' sector are used. In theory, the vector mean chart gives an accurate picture of the net flow of water in the long run, but in practice its accuracy, and that of the predominant chart, depend upon the number of observations available. This number is not yet adequate in many parts of the oceans, even for quarterly charts.

Current Atlases Issued by the Meteorological Office

The following are available (1966):

(*a*) Met. O. 485. *Quarterly Surface Current Charts of the Western North Pacific, with Monthly Chartlets of the China Seas* (1965).

(*b*) Met. O. 466. *Quarterly Surface Current Charts of the Atlantic Ocean* (1962).

(*c*) Met. O. 655. *Quarterly Surface Current Charts of the Eastern North Pacific* (1960).

(*d*) Met. O. 772. *Indian Ocean Currents* (1939).

(*e*) Met. O. 435. *South Pacific Ocean Currents* (1939).

The South Pacific atlas is now being revised. In general, each atlas is revised when enough new current observations have been received. At present (1966) only the first three of the above atlases contain predominant charts; these will be included in all future editions.

CHAPTER 15

THE CURRENT CIRCULATIONS OF THE OCEANS

Introduction

A summary of the general surface current circulation of all oceans is given below. The general circulation represents the trend of water movement in the long run, emerging from the more or less variable movements of individual currents.

In each ocean the main circulation is dealt with first, followed by any extensions thereof (associated circulations). The surface circulation of the Mediterranean, for example, is not a part of the main North Atlantic circulation, but is associated with it.

The general surface current circulation of the world is shown in the World Current Chart, opposite page 202, on which the variations during the two monsoon seasons are shown. Apart from these major changes of direction, there are some minor seasonal changes in position of currents, which cannot be shown on a single chart. One of these is the position of origin of the Equatorial Countercurrent in the North Atlantic, referred to later.

The fullest information available will be found in the latest editions of the current atlases, a list of which is given on page 200. The more information received from mariners, the more accurate the atlases will be.

Information about the currents in local coastal regions, so far as these are known, will be found in the latest editions of the *Admiralty Pilots*. This is based on the current atlases, together with all available local knowledge. Current information is also being supplied for many Admiralty navigational charts as these are revised.

NORTH ATLANTIC OCEAN

Main Circulation

The main circulation of the North Atlantic is clockwise. The southern part of this consists of a wide belt of west-going current, comprising the fairly constant NORTH EQUATORIAL CURRENT, southward of about lat. 20°N, and the less constant NORTH SUBTROPICAL CURRENT, between about lat. 20°N and 32°N. Eastward of the Caribbean Sea, the North Equatorial Current is joined by the SOUTH EQUATORIAL CURRENT of the South Atlantic, which flows past the north coast of Brazil. The combined Equatorial Current flows westward through the Caribbean Sea and emerges through Yucatan Channel. It thence flows north-eastward along the north-west coast of Cuba and into Florida Strait as the beginning of the GULF STREAM, which forms the west side of the circulation.

The North Subtropical Current, also flowing westward, passes into, or turns north and runs parallel to, the seaward edge of the Gulf Stream, in all latitudes from the north coast of Cuba to about lat. 32°N. That part of the current which passes through the Bahamas is known as the BAHAMA or ANTILLES CURRENT.

In a narrow belt north of the equator, the EQUATORIAL COUNTERCURRENT flows eastward, between the North Equatorial Current of the North Atlantic and the South Equatorial Current of the South Atlantic. Its longitude of origin varies, being about 52°w in August to October, and 18°w to 26°w in February to April. This countercurrent flows past Cape Palmas and sets along the coast of the Gulf of Guinea to the Bight of Biafra, where it is known as the GUINEA CURRENT. The subsequent course of the Guinea Current is not known; most of it probably recurves westward and joins the South Equatorial Current.

The warm waters of the Gulf Stream, between Florida Strait and Cape Hatteras, follow the course of the 100-fathom line, outside but near which its axis of greatest strength lies. Immediately north of Cape Hatteras the Gulf Stream begins to leave the 100-fathom line and gradually turns eastward into the ocean, southward of the Georges and Nova Scotian Banks. The inshore edge of the Gulf Stream, south of Cape Hatteras, thus gradually becomes the northern edge, northward and eastward of that cape. This northern edge is relatively sharply defined at all times of the year, owing to the convergence along it of the cold Labrador Current (*see* below).

Eastward of about the 46th meridian the Gulf Stream ceases to be a well-defined current. It weakens by fanning out up the east side of the Great Bank of Newfoundland. The resulting north-east and easterly flow across the ocean is towards the British Isles and adjacent European coasts; this is known as the NORTH ATLANTIC CURRENT.

The southerly part of the North Atlantic Current turns gradually clockwise to south-easterly and then south-westerly directions, east of the 40th or 45th meridians. It thus passes into the North Subtropical Current to complete the main circulation. This wide southerly flow, called the AZORES CURRENT, occupies the belt between about the 42nd and 32nd parallels. The PORTUGAL CURRENT, flowing southward off the west coasts of Spain and Portugal, and the CANARY CURRENT further southward, form the coastal fringe of the Azores Current. The Canary Current passes into the North Equatorial Current between the latitudes of Cape Blanco and Cape Verde.

Northward Extension of the Main Circulation

The more northerly part of the North Atlantic Current does not recurve southward, but flows north-eastward off the west coasts of the Hebrides and Shetland Islands, and thence to the coast of Norway. It sets northward up this coast, where it is known as the NORWEGIAN ATLANTIC CURRENT. In about lat. 69°N this current divides, and the left branch, the WEST SPITSBERGEN CURRENT, sets northward to the west coast of Vestspitsbergen and thence into the Arctic Basin. The right branch is called the NORTH CAPE CURRENT and follows the coast past North Cape into the Barents Sea, finally setting towards the north of Nóvaya Zemlyá; a branch of it continues along the Múrmanski coast as the MÚRMANSKI CURRENT.

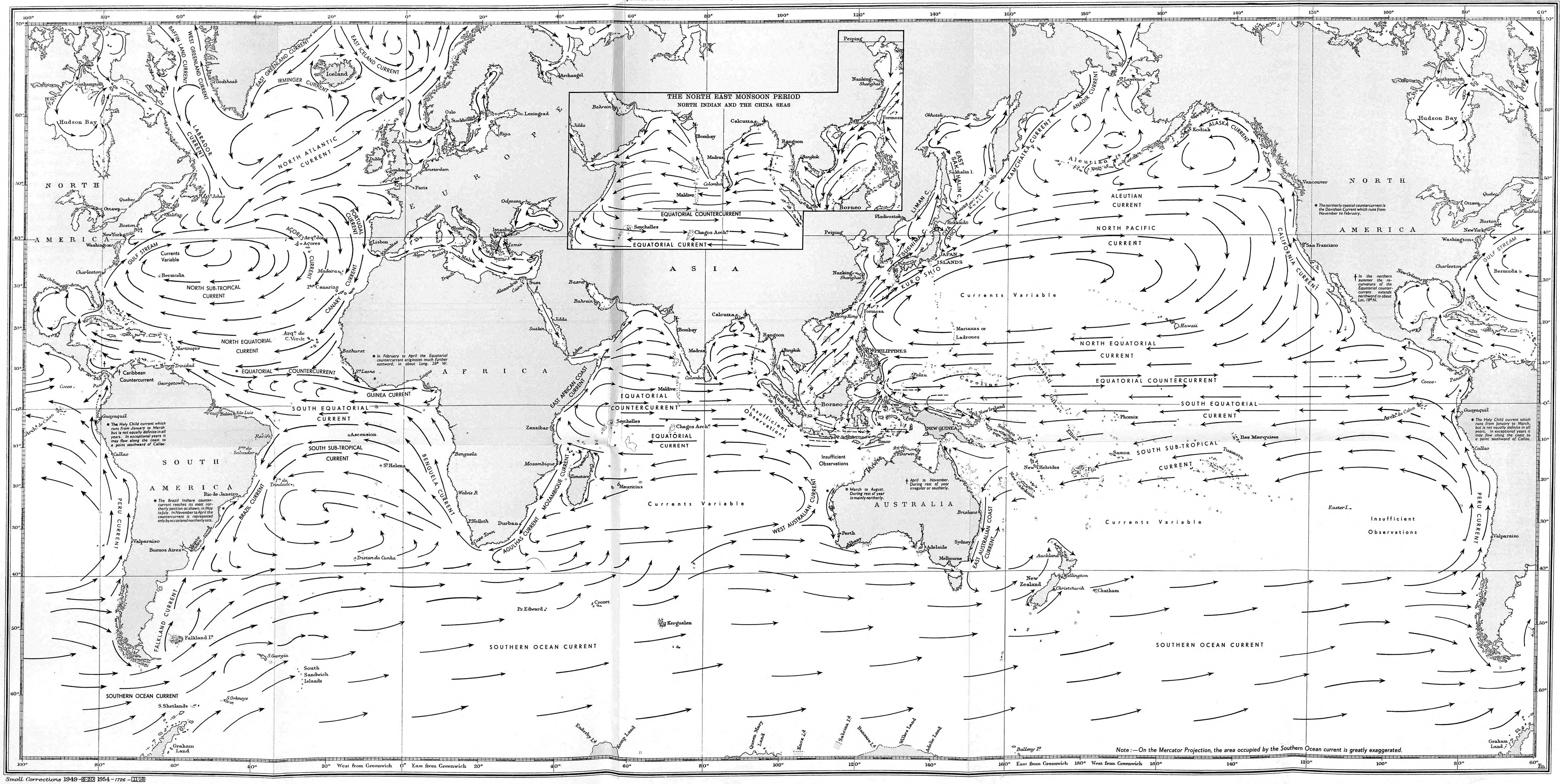
The branch of the North Cape Current which flows into the Barents Sea towards and round the northern extremity of Nóvaya Zemlyá, continues south-westward in the Kara Sea along the east coast of Nóvaya Zemlyá, where it is known as the NÓVAYA ZEMLYÁ CURRENT. Part of this re-enters the Barents Sea along the northern shore of Kara Strait, forming the LITKE CURRENT.

A small part of the warm North Atlantic Current turns northward in the longitude of Iceland to form the IRMINGER CURRENT. Closely south-west of Iceland it divides, and the main branch turns westward and passes into the East Greenland Current south of Denmark Strait. A smaller branch makes a clockwise circulation of Iceland.

THE WORLD

GENERAL SURFACE CURRENT CIRCULATION

The circulation of the south-west monsoon period in the North Indian Ocean and China Seas is shown on the main chart. The circulation of the north-east monsoon period in these areas is given on the inset chartlet. In the Eastern Archipelago and the region north of New Guinea the circulation of both monsoon periods is shown on the main chart, that of the north-east monsoon period being indicated by pecked lines



Small Corrections 1949–6.20 1954–1726–11.16

Fig. 15.1. General surface current circulation of the world

The chief outflow of water from the Arctic Basin is the cold, ice-bearing current which sets south-westward along the east coast of Greenland, the EAST GREENLAND CURRENT. A part of this diverges south-eastward from the main body of the current, north of lat. 70°N. This, the EAST ICELAND CURRENT, passes the north-east coast of Iceland and north of the Faeröes, gradually trending eastward and finally north-eastward. It joins, or runs parallel to, the outer edge of the Norwegian Atlantic Current.

The East Greenland Current rounds Cape Farewell and passes northwards along the west coast, where it is called the WEST GREENLAND CURRENT. This loses volume by fanning out on its seaward side, but part of it circuits the head of Baffin Bay, and, reinforced by water flowing eastward through Jones and Lancaster Sounds, sets southward along the coast of Baffin Island as the BAFFIN LAND CURRENT. Northward of Hudson Strait this is joined by another branch from the West Greenland Current crossing Davis Strait. The combined current, known as the LABRADOR CURRENT, sets past Hudson Strait entrance and south-eastward along the Labrador coast towards Newfoundland.

After passing Belle Isle Strait and the east coast of Newfoundland, the Labrador Current covers the whole of the Great Bank except, during the summer, the extreme southern part. A large branch of the current follows the eastern edge of the Bank, thus carrying ice furthest south to reach the trans-atlantic steamship tracks. Another branch rounds Cape Race, and sets south-westerly. The bulk of the water on the Bank also sets in this direction, so that the Labrador Current fills the region between the south coast of Newfoundland, the south-east coast of Nova Scotia and the northern edge of the Gulf Stream. The Labrador Current continues southward as a cold current along the United States coast. Its greatest southern extension is to about lat. 36°N, closely north of Cape Hatteras, in November to January; its least extension is to about lat. 40°N in August to October. The cold Labrador Current and the warm Gulf Stream converge along the northern edge of the latter.

Water also emerges from the Arctic Ocean into the northern part of the Barents Sea, forming the EAST SPITSBERGEN CURRENT, which flows south-westward and passes the southern extremity of Vestspitsbergen, where it curves north-westward into the West Spitsbergen Current. A similar current, further south, is directed towards Bear Island, and is known as the BEAR ISLAND CURRENT. Part of this recurves northward into the West Spitsbergen Current and part southward.

Associated Regions

North Sea and English Channel

This region is one of tidal streams, and variable currents depending on the present or recent wind. After subtracting the tidal streams, there is in the long run a weak current circulation as follows. North-eastward of the Shetland Isles a branch of the North Atlantic Current flows southward down the east coasts of Scotland and England to the Thames estuary, where it recurves to the eastward. It is there joined by a branch of the North Atlantic Current which passes up the English Channel and through the Strait of Dover. The combined current flows along the Belgian and Netherlands coasts and west coast of Jutland. It continues counterclockwise round the Skaggerak and then sets up the west coast of Norway.

Only a portion of the current on the west side of the North Sea reaches as far south as the Thames estuary, for water fans out eastward from this current all along its length. The bulk of this water, after crossing the North Sea, enters the Skaggerak on its southern side and so joins the current up the Norwegian coast.

The current flowing northward up the west coast of Norway is probably the most constant one in the North Sea area. It is the outflow from the North Sea, and in about lat. 62°N it rejoins the main branch of the North Atlantic Current, the combined current flowing northward towards the North Cape. The outer part of this current is called the NORWEGIAN ATLANTIC CURRENT, as previously stated, while the inner part, which follows the coastline, is known as the NORWEGIAN COASTAL CURRENT.

Bay of Biscay

Off the mouth of the Bay of Biscay the current trends south-eastward and southward to form the beginning of the Portugal Current. A branch enters the Bay and recurves westward along the north coast of Spain to rejoin this current near Cape Finisterre.

Mediterranean

Part of the water from the Portugal Current enters the Strait of Gibraltar and flows along the north coast of Africa. Passing Cape Bon it continues in a general south-easterly direction towards Port Said. Turning northward at the eastern end of the Mediterranean, the counterclockwise circulation is completed by a more variable return current along the northern coasts. In following the coast, this forms counterclockwise loops in seas such as the Aegean and the Adriatic. The outflow from the Mediterranean is a subsurface one, westward through the Strait of Gibraltar, beneath the inflowing surface current. (*See* page 224.)

Black Sea

The general circulation is counterclockwise. There is an almost constant surface flow of water from the Black Sea to the Aegean through the Bosporus, Sea of Marmara and the Dardanelles. There is a subsurface return current below this, from the Aegean to the Black Sea.

Gulf of St. Lawrence

Water enters the Gulf on the northern side of Cabot Strait and flows up the west coast of Newfoundland. It then turns south-westward along the Quebec coast and is west-going southward of Anticosti. The GASPÉ CURRENT sets south-eastward from the Gaspé Peninsula. Water emerges from the Strait by the CAPE BRETON CURRENT on the southern side of Cabot Strait; this is sometimes called the CABOT CURRENT.

Gulf of Mexico

Part of the water passing through Yucatan Channel turns westward and follows the gulf coast in a clockwise direction. Another branch sets northward across the middle of the gulf to the region of the Mississippi delta, where it turns eastward and joins the coastal current. The combined current passes into the Gulf Stream between Cuba and Florida.

Caribbean Countercurrent

This sets eastward along the coasts of Panama and Colombia to the Rio

Magdalena and beyond, throughout the year. It is strongest in August to October.

SOUTH ATLANTIC OCEAN

Main Circulation

The main surface circulation of this ocean is anticlockwise. The SOUTH EQUATORIAL CURRENT, flowing westward across the ocean, extends across the equator to about lat. 4°N; southward of about lat. 6°S there is a big reduction in the average strength of the current. Between lat. 6°S and 22°S the weaker general westerly flow is known as the SOUTH SUBTROPICAL CURRENT.

The eastern side of the circulation is formed by the relatively cool BENGUELA CURRENT, flowing north-westerly from the region of Cape Town and then into the South Equatorial Current. From about lat. 30°S, water from the Benguela Current fans out west-north-westerly and westerly on its seaward side. Northward of lat. 22°S this water passes into the South Subtropical Current.

While most of the South Equatorial Current flows along the north coast of Brazil and across the Equator to join the North Equatorial Current, the South Subtropical Current is directed towards the Brazilian coast, southward of Cape St. Roque. A small part of this, turning northward along that coast, passes round Cape St. Roque and joins the South Equatorial Current; the bulk of it flows southward along the Brazilian coast. This warm current, known as the BRAZIL CURRENT, forms the west side of the South Atlantic circulation.

The SOUTHERN OCEAN CURRENT flows all round the globe southward of the South Atlantic, South Indian and South Pacific Oceans, and forms the completion, on the southern side, of the anticlockwise circulation in these oceans. The Southern Ocean Current is restricted in width by passing through Drake Passage, between Cape Horn and Graham Land. Eastward of this passage it becomes very wide, its northern part fanning out north-eastward past the southern and eastern coasts of the Falkland Islands, to reach to about the 38th parallel in the central longitudes of the South Atlantic.

Further north, the southern part of the main circulation is added to by water from the seaward side of the Brazil Current recurving south-eastward and eastward between lat. 28°S and 42°S. In mid-ocean part of the resulting easterly flow runs north of and parallel to the colder water of the Southern Ocean Current; the remainder merges with the northerly part of the Southern Ocean Current. East of about long. 15°W and south of the 30th parallel, the east-going water turns north-east and north to converge with the westerly flow fanning out from the seaward side of the Benguela Current.

Nearer the South African coast, between long. 10°E and 15°E, a branch of the Southern Ocean Current turns northward directly into the Benguela Current. The water of the Benguela Current is, however, mainly derived by the upwelling of water from subsurface depths off the south-west coast of Africa. A branch of the AGULHAS CURRENT of the South Indian Ocean, which rounds the south coast of Africa, also enters the Benguela Current.

The FALKLAND CURRENT does not form part of the main circulation; it branches northward from the Southern Ocean Current near Staten Island and passes west of the Falkland Islands. Part of it continues to the Rio de la Plata estuary; the remainder branches eastward in about lat. 40°S to 42°S and rejoins the northern part of the Southern Ocean Current. During May to October, a northerly extension of the Falkland Current known as the BRAZIL INSHORE

COUNTERCURRENT continues north of Rio de la Plata. From May to July this may extend as far as Cape Frio.

NORTH INDIAN OCEAN

Monsoonal Effects

The currents in the North Indian Ocean are reversed in direction seasonally by the monsoons. These comprise the currents of the Arabian Sea and Bay of Bengal and the EAST AFRICAN COAST CURRENT,* between lat. 4°N and Cape Guardafui. The only current north of the equator which is not reversed in direction is the EQUATORIAL COUNTERCURRENT, which lies mainly south of the equator but extends a few degrees north of it.

The NE monsoon circulation occurs during the height of that monsoon in November to January. During the latter part of this monsoon, February to April, the circulation changes. The SW monsoon circulation prevails from May to September. October is a transition month. The currents are therefore described below for the three periods, November to January, February to April and May to September.

North-east monsoon circulation, November to January. In the open waters of the Arabian Sea and Bay of Bengal the current sets in a westerly direction. Owing to the coastal conformation, it flows counterclockwise round the coasts. A stronger current sets southward down the east coast of Africa from Ras Hafun to about lat. 2°S. This, the EAST AFRICAN COAST CURRENT, turns eastward between about lat. 2°N and 2°S, to form the beginning of the Equatorial Countercurrent. Between Ras Hafun and Cape Guardafui the current is northerly.

Later north-east monsoon period, February to April. The flow in the open waters of the Arabian Sea and Bay of Bengal remains westerly, though the currents are more variable than in November to January. The coastal circulation, both of the Arabian Sea and Bay of Bengal is, however, reversed to a clockwise direction (*see* page 221). This reversal is completed in the Bay of Bengal by about the beginning of February. In the Arabian Sea it is more gradual and is not complete on all parts of the coast until the end of March. The direction of the East African Coast Current is in transition in March and flows northward along the coast in April, from Cape Delgado to Cape Guardafui.

South-west monsoon circulation, May to September. The coastal circulation of the Arabian Sea and Bay of Bengal remains clockwise and is strengthened. The East African Coast Current also continues to flow northward, from Cape Delgado to Cape Guardafui, and is greatly strengthened. It divides in about lat. 7°N; part continues along the coast to Cape Guardafui, but the bulk turns eastward and passes south of Socotra into the general easterly current. The current south of Socotra in July to September is the strongest known in the world in the open ocean, rates up to 7 knots having been recorded. In the open waters of the Arabian Sea and Bay of Bengal the current sets to the east.

Equatorial Countercurrent

The origin of the Equatorial Countercurrent is not the same throughout the year. In the earlier part of the NE monsoon, November to January, it arises from the south-going East African Coast Current. Current observations are not numerous enough to determine the northern limit of this current in all longitudes. In November to January the easterly flow extends to lat. 2°N, west of

* This current is often called the SOMALI CURRENT.

long. 80°E, but between that meridian and the coast of Sumatra it appears to extend to lat. 4°N.

In February to April the Countercurrent originates west of the Seychelles, by a recurvature southward of the general westerly flow in the southern Arabian Sea. South-westward of the Seychelles a branch of the Equatorial Current turns northward to join the Countercurrent. West of long. 80°E, the northern limit of the Countercurrent appears to be south of the equator, but east of that meridian it reaches the equator, or lat. 2°N.

During the SW monsoon, May to September, the Countercurrent flows in the same direction as the easterly monsoon current of the Arabian Sea and Bay of Bengal. In the western part of the ocean it is distinguished by the current south of lat. 2°N being usually stronger than that to the northward.

The course of the Countercurrent on the east side of the ocean is not fully known. In November to January the bulk of it appears to follow the west coast of Sumatra and the south coast of Java, in south-easterly and easterly directions. During the SW monsoon it seems that part of it recurves southward, west of Sumatra, passing into the Equatorial Current.

Red Sea and Gulf of Aden

The current conforms to the monsoon. During the NE monsoon it sets westward in the Gulf of Aden and, passing through the Straits of Bāb-el-Mandeb, flows up the axis of the Red Sea. During the SW monsoon the current in the Gulf of Aden sets eastward; in the Red Sea the water flows down the axis of the sea and into the Gulf of Aden.

SOUTH INDIAN OCEAN

Main Circulation

The main circulation of the South Indian Ocean is counterclockwise. Owing to the monsoons there is only one Equatorial Current in the Indian Ocean, and this corresponds to the South Equatorial Current of the Atlantic or Pacific. The west-going flow of the Indian Ocean EQUATORIAL CURRENT lies well south of the equator, thus differing from those of the Atlantic and Pacific, which extend to a few degrees north of the equator. Its northern boundary is usually between lat. 6°S and 10°S, varying according to longitude and season.

The west-going Equatorial Current, after passing the northern extremity of Madagascar, meets the African coast near Cape Delgado. Here it divides, some of the water flowing northward up the coast. The remainder flows southward, forming a strong coastal current, which from Cape Delgado to Delagoa Bay is known as the MOZAMBIQUE CURRENT. Its southward continuation is the AGULHAS CURRENT. This is reinforced by water from the Equatorial Current setting past the southern extremity of Madagascar.

Some of the water of the Agulhas Current recurves to south-eastward between about long. 20°E and 32°E and enters the northern part of the Southern Ocean Current. The remainder of the Agulhas Current continues along the coastline and, passing over the Agulhas Bank, enters the South Atlantic Ocean, where it joins the Benguela Current.

The southern side of the main circulation is formed by the cold water of the SOUTHERN OCEAN CURRENT, setting easterly to north-easterly as far as long. 80°E, and easterly to south-easterly on the east side of the ocean. As above stated, some of the warm Agulhas Current water also contributes to this part of the circulation. There is no defined northern boundary to the Southern Ocean Current; the predominance of easterly sets decreases with decreasing latitude towards the middle of the ocean, until it merges into the region of variable current south of the Equatorial Current. Some predominance of easterly set is found as far north as lat. 28°S or 30°S in mid-ocean.

The east side of the circulation is not well marked. The Southern Ocean Current on approaching Australia sends off a branch which passes into the WEST AUSTRALIAN CURRENT, a weak north-westerly flow off the west coast. This passes into the Equatorial Current in about lat. 16°S to 20°S, long. 95°E to 105°E. The bulk of the Southern Ocean Current continues its easterly course, south of Australia and Tasmania, into the South Pacific.

The EQUATORIAL COUNTERCURRENT flows eastward throughout the year, immediately north of the Equatorial Current. It lies mainly south of the equator but during the NE monsoon season it also extends northward. It is more directly connected with the North Indian Ocean currents than with those of the South, and is more fully described on page 206.

Extreme Eastern Part of the Ocean

The currents here, including those of the Arafura Sea, are not well known, owing to scarcity of observations. Eastward of Christmas Island, between the parallels of about 10°S and 12°S, there is a predominance of westerly sets, during most of the year, forming the most easterly part of the Equatorial Current.

NORTH PACIFIC OCEAN

Main Circulation

The main circulation of the North Pacific resembles that of the North Atlantic. Observations are inadequate to give detailed information about the currents over large parts of this ocean, owing to its size and limited shipping tracks. This especially applies to the middle longitudes, both near the equator and in the variable current region further north.

The southern part of the main circulation is formed by the west-going NORTH EQUATORIAL CURRENT. Immediately south of this the EQUATORIAL COUNTERCURRENT flows eastward across the ocean, but its limits are not exactly known and may be subject to some seasonal variation. During the latter half of the year the southern limit appears to be nearer the equator in the west than in central or eastern longitudes. Over most of the ocean the countercurrent is usually found between lat. 4° or 5°N and 8°N. The SOUTH EQUATORIAL CURRENT, the northern limit of which reaches to about lat. 4°N, is described under the South Pacific.

The North Equatorial Current has no defined northern limit. This west-going current lessens in strength as the predominance of trade winds decreases, until it is lost in the variable current region lying to the northward. The latitude to which some predominance of westerly current extends appears to vary with the season. In mid-ocean it is between lat. 20°N and 24°N in winter and about lat. 30°N in the late summer or autumn.

The Equatorial Countercurrent flows eastward throughout the year across the whole ocean. During March to November, this countercurrent is formed by the recurving of the South Equatorial Current northward and part of the North

Equatorial Current southward down the east coasts of the Philippines. In December to February the North Equatorial Current is the only source of the countercurrent. During these months the South Equatorial Current, north of the equator, turns south in about long. 140°E to 150°E and finally south-eastward, and thus plays no part in forming the countercurrent. In all seasons part of the North Equatorial Current enters the Celebes Sea, emerging therefrom in a north-easterly direction to join the countercurrent. The countercurrent appears to be strongest in its most westerly portion, from northward of Halmahera (between New Guinea and Celebes) to about long. 145°E.

To continue the main circulation, a considerable part of the water from the North Equatorial Current turns north-eastward when east of Luzon and flows up the east coast of Formosa to form the KURO SHIO, a warm current corresponding to the Gulf Stream of the North Atlantic. Southward of the Japanese Islands the Kuro Shio flows north-east. This current then fans out to form the NORTH PACIFIC CURRENT, which sets eastward across the ocean to the American coast. It is joined by cold water from the Bering Sea, flowing down the east coast of Kamchatka and turning south-east and then east. The whole forms a broad belt of variable current with a predominance of easterly set, filling most of the area between lat. 35°N and 50°N, across the ocean. The colder part of this is known as the ALEUTIAN or SUBARCTIC CURRENT and in the middle longitudes of the ocean is found northward of about lat. 42°N.

Water fans out south-east and south from the southern part of the North Pacific Current, and passes into the variable current region. Eastward of about long. 160°W the remainder of the North Pacific Current and the bulk of the Aleutian Current turn south and south-westward, finally passing into the North Equatorial Current. Near the coast this is called the CALIFORNIA CURRENT.

The California Current does not actually meet the coast; from November to February a relatively cool countercurrent, known as the DAVIDSON CURRENT, runs northward, close inshore, to at least lat. 48°N. During the rest of the year the space between the California Current and the coast is filled by irregular current eddies. (*See* page 96 for the frequency of fog in this area.)

In the extreme eastern part of the Equatorial Countercurrent, seasonal variations occur off the Central American coast and numerous eddies are formed, which seem to vary from year to year. In most months the countercurrent will be met between lat. 5°N and 6°N, and it generally turns north and north-west along the Central American coast, finally entering the North Equatorial Current. Early in the year part of the Countercurrent branches south and enters the South Pacific (*see* page 211). There is an inflow during most of the year into the Gulf of Panama, on its west side from the Countercurrent, and on its east side from the Peru Current; the resultant southerly outflow in the middle of the Gulf crosses the meridian of 80°W and turns south-west to enter the South Equatorial Current.

Associated Regions

Northern Part of the Ocean

The Bering Sea currents are not well known, but there is a counterclockwise circulation round the coasts, northward on the east and southward on the west side. This cold southward current flows along the east coast of Kamchatka as the KAMCHATKA CURRENT, and then past Chisima Rettō, where it is known as the

OYA SHIO.* The Oya Shio continues along the east coast of the main Japanese island of Honshu, until it meets the northern edge of the Kuro Shio in about lat. 36°N. The Oya Shio thus corresponds to the Labrador Current of the North Atlantic. (*See* page 96 for the frequency of fog in this area.) It is joined by water emerging through Tsugaru Kaikyō. Water fans out south-east and east all along the course of the current, from Kamchatka southward. The resulting easterly current flows parallel and adjacent to the North Pacific Current, forming the Aleutian Current, the bulk of which, as already stated, sets southward, on the east side of the ocean, as the California Current. The remainder sets north-east and then north-west past Queen Charlotte Islands and along the coast of south-east Alaska. This current, the ALASKA CURRENT, is reinforced, during November to January, by water from the Davidson Current, which sets up the American coast as far as Vancouver Island. The Alaska Current follows the Gulf of Alaska coastline, setting to the westward across its head and then along the south coasts of the Aleutian Islands. West of long. 155°W to 160°W, some water recurves from the Alaska Current to the south and south-east and rejoins the east-going Aleutian Current; the remainder recurves northward and enters the Bering Sea; thence turning north-east and eastward it forms the east side of the Bering Sea circulation, referred to above.

China Seas and other Regions Westward of the Main Circulation

In the China Seas and Java Sea the currents are monsoonal. During the SW monsoon the general direction of current is west in the Java Sea, north-east in the China and Eastern Seas and north in the Yellow Sea. During the NE monsoon these directions are reversed, becoming south in the Yellow Sea, south-west in the Eastern and China Seas and east in the Java Sea. In the southern part of the China Sea there is a variable current area west of Borneo and Palawan in both monsoons, but a weak monsoonal current runs along the west coasts of these islands, alternating between north-east and south-west during the year. In the east part of the Eastern Sea the Kuro Shio flows.

In the China Seas, the north-east current is found from May to August inclusive. September is the transition month, but the north-east current still persists in the southern part. In October the south-west current becomes established everywhere, and this continues till about the middle of March. April is the transition month, but the south-west current still persists in the southern part.

In the Java Sea the westerly current runs from June to September and the easterly current from November to March. April, May and October are transition months.

In the Japan Sea the circulation is counterclockwise all the year, the north-going current on the east side of the sea, known as the TSUSHIMA SHIO, being a branch of the Kuro Shio which has passed through Korea Strait. Part of the Tsushima Shio branches off through Tsugaru Kaikyō and flows into the Oya Shio, and another part branches off through Sōya Kaikyō. The south-going current on the west side of the sea, past Vladivostok, is called the LIMAN CURRENT.

There is little or no current in the central part of the Sea of Okhotsk. A counterclockwise current flows round the coast. Off the east coast of Sakhalin Island the south-going current is known as the EAST SAKHALIN CURRENT.

* The whole of the cold current from the Bering Sea is sometimes referred to as the Oya Shio, especially by American writers.

SOUTH PACIFIC OCEAN

Main Circulation

The main circulation is counterclockwise. Less is known about the South Pacific currents than those of the other oceans south of the equator, because of its great size and on account of the large areas, particularly in the east, which are not frequented by shipping.

The SOUTH EQUATORIAL CURRENT of the Pacific, though lying mainly south of the equator, extends from 1° to 4° or 5°N in different longitudes and seasons. Its northern limit is defined by the east-going Equatorial Countercurrent of the North Pacific, immediately north of it.

South of about lat. 6°S the average strength of this current is much reduced, though its westerly direction remains. Between about lat. 6°S and 20°S this weaker and less constant westerly current is known as the SOUTH SUBTROPICAL CURRENT.

On the west side of the ocean the course of the South Equatorial Current varies seasonally. In June to August the whole current follows the north coast of New Guinea to the north-westward and then recurves north and north-eastward, entering the east-going Equatorial Countercurrent of the North Pacific; also in September to November and March to May some water recurves from the northern part to join this Countercurrent. In December to February the South Equatorial Current does not pass into the Countercurrent; it recurves south-west and southward and flows past the north coast of New Guinea in a south-easterly direction. There is thus a complete reversal of current along this coast during the year.

The west side of the main circulation is not well marked, except along the Australian coast from Great Sandy Island to Cape Howe, where the EAST AUSTRALIAN COAST CURRENT sets southward. This is not a very constant current, and northerly sets may occur off this coast at any time of the year.

Water flows south-westward, in the long run, from the South Subtropical Current to enter the East Australian Coast Current, past Ellice Islands, the new Hebrides and New Caledonia, but the currents in this region, so far as they are known, show much variation. Part of the South Subtropical Current flows through Torres Strait into the Indian Ocean. Little is known of the Coral Sea currents except that in the north the set is towards Torres Strait and in the south it is south-westerly or southerly, towards the East Australian Coast Current.

The south side of the main circulation is formed by the SOUTHERN OCEAN CURRENT, setting easterly or north-easterly. Observations are scanty over most of this region, but the current seems to be weaker and more variable than in the South Atlantic and Indian Oceans, and northerly sets appear not infrequent.

Between Australia and New Zealand the current is variable with some predominance of easterly sets. The bulk of the East Australian Coast Current mixes with the water of the Southern Ocean Current flowing eastward south of Tasmania and also through Bass Strait; part of this flows north-easterly along the west and east coasts of South Island, New Zealand.

The bulk of the Southern Ocean Current enters the South Atlantic south of Cape Horn. The northern part of this current, however, meets the coast of Chile between Isla Chiloe and the Golfo de Penas, where it divides, part going northward to form the beginning of the Peru Current and part following the coast south-eastward to rejoin the Southern Ocean Current south of Cape Horn.

The east side of the circulation is formed by the relatively cool PERU CURRENT, formerly known as the HUMBOLDT CURRENT. It follows the coastline northward to the equator. Between the Gulf of Guayaquil and the equator the bulk of the Peru Current trends seaward and passes into the South Equatorial Current. The Peru Current has a width of perhaps 300 miles or more.

A branch of the Peru Current continues northward up the coast during most of the year and enters the Gulf of Panama.

During the northern winter in the eastern North Pacific, the east-going Equatorial Countercurrent extends further south than at other seasons. A branch of this current then turns southward along the coast of Ecuador into the South Pacific, but in most years its southern limit is only a few degrees south of the equator. This is called EL NINO, or the HOLY CHILD CURRENT, as in some years it begins to flow about Christmas-time, although it is more regularly observed in February and March. In exceptional years it extends down the coast of Peru, occasionally to beyond Callao, when the intrusion of this warm water into a region normally occupied by the cool Peru Current kills fish and other marine life. (*See* page 276.)

Central Oceanic Region

Because of the great width of the South Pacific there is in the centre, between about the 20th and 45th parallels, the largest area of variable current in the world. Current observations in this area are scanty, particularly on its eastern side. No general trend of current is shown during any part of the year. Between New Zealand, Norfolk Island, Fiji and Tonga, while currents in all directions may occur, there is some predominance of flow between north and east, particularly from May to October.

ARCTIC OCEAN

The main inflow of water into the Arctic Basin is from the West Spitsbergen Current. A much smaller quantity enters through Bering Strait. Fresh water is added to the Arctic Basin from rivers, notably those of Siberia, and by an excess of precipitation over evaporation.

The EAST GREENLAND CURRENT forms the main outflow of water from the Arctic Ocean. Small outflows occur due to the EAST SPITSBERGEN and BEAR ISLAND CURRENTS, flowing south-westward in the northern part of the Barents Sea, and the current flowing eastward between the islands of the Arctic Archipelago towards Baffin Bay.

Within the eastern longitudes of the Arctic Basin there is a weak westerly current, as shown by the drift of the *Fram* and other ships in the ice. This emerges to the south-west between Svalbard (Spitsbergen) and Greenland to form the East Greenland Current.

Along the Siberian coast there is a general current to the eastward from the Kara Sea to Bering Strait.

ANTARCTIC OCEAN

The east-going SOUTHERN OCEAN CURRENT completes, on the south side, the anticlockwise circulation in the South Atlantic, South Indian and South Pacific Oceans. The southern limit of this current lies in about lat. 66°S in the Atlantic and Indian Oceans; in the South Pacific there is still an easterly set south of this latitude, as the westerly winds extend further south.

Near the Antarctic coastline a relatively strong westerly current follows the trend of the coast. This is known as the ANTARCTIC COASTAL CURRENT. The current is produced by the general easterly winds of these very high latitudes. It begins in the Bellinghausen Sea and, after circumnavigating the continent to the Weddell Sea, is diverted north-eastward by the Graham Land peninsula, whence it enters the Southern Ocean Current.

CHAPTER 16

THE CAUSES AND CHARACTERISTICS OF OCEAN CURRENTS

Introduction

The water of the oceans is in a state of continual movement, not only at the surface but at all depths. Ocean current circulation, in its widest sense, takes place in three dimensions, but the stronger currents occur in an upper layer which is shallow compared with the ocean depth. A current at any depth may flow horizontally, obliquely or vertically, although a surface one only flows horizontally.

The navigator is interested only in the surface current, extending to a depth of about half the ship's draught. This may differ slightly from that close to the surface, such as would affect a boat.

The processes which cause ocean currents are complex and not yet fully understood. In most cases, it is probable that more than one factor is responsible for a particular current at a given time. There is no geographical beginning of most surface currents, as the main currents in each ocean form a closed circulation. Nor is there any beginning in time of the general circulation as a whole, which must have been essentially the same since land and oceans took their present form. There is thus an endless chain of cause and effect at work, stretching back into the past. A current produced by a wind, for example, may itself become a cause of some subsequent water movement, owing to the resulting disturbance of the equilibrium of the sea.

General Surface Circulation

If all the currents observed within a small area of the ocean be plotted, it will be found that they are variable, to some extent. Over a period of time surface water thus flows into and out of the area in various directions. Each individual current is never exactly balanced by one of the same strength in the opposite direction; there will thus be a resultant flow of water across the area in the long run. This is found by working out the vector mean of all the currents, i.e. taking account of the direction as well as rate. These flows across all areas of the ocean make up the general circulation, which is the general pattern of the currents, after eliminating local day-to-day variations.

The primary cause of surface currents in the open ocean is the direct action of wind on the sea surface; thus there is a close relation between the directions of currents and prevailing winds (*see* Figs. 16.1 and 16.2). A constant wind blowing over great stretches of an ocean will have the greatest effect; thus the north-east and south-east trade winds are the mainspring of the surface current circulation. In the Atlantic and Pacific the trade winds drive an immense body of water westward over a width of some 50° of latitude, broken only by the narrow belt of east-going Equatorial Countercurrent. The south-east trade produces a similar westerly current in the South Indian Ocean.

The main surface current circulations thus form 'closed' systems or eddies in accordance with the winds around the permanent anticyclones centred about latitudes 30°N and S (*see* Chapter 10). The direction of circulation is clockwise in the northern and anticlockwise in the southern hemisphere. There are also current circulations outside the main eddies. Thus part of the North Atlantic Current branches from the main system and flows north of Scotland and along the coast of Norway. The current circulation of the Mediterranean, though an entity in itself, is associated with that of the Atlantic.

In the monsoon regions the current reverses seasonally, according to the monsoon.

The Southern Ocean Current, produced by the westerly winds of the Roaring Forties, encircles the globe in an easterly direction, to form the southern part of the circulation in the South Atlantic, South Indian and South Pacific Oceans.

Current Variability

In all oceans the currents in a given region are variable; this is mainly due to changes of wind, in strength and direction. Even in the more variable regions there is often, however, a greater frequency of current setting towards one direction, so that there is an eventual flow of water across the area in the direction of the general circulation. Some variability, including occasional currents opposite to the usual flow, occurs even in the more constant currents, such as the Gulf Stream. The constancy of such currents varies in different seasons and in different parts of the current. It is usually about 75%, rarely exceeding 85%, and this only in very limited areas.

Current roses fall into three main types. Fig. 16.3*a* shows a typical rose of a region of strong and fairly constant current. Fig. 16.3*b* shows a typical rose of a monsoon or trade-wind area, where predominance of direction is not so marked as in Fig. 16.3*a*. Fig. 16.3*c* shows a rose of a 'variable current' area, the frequency being nearly equal in all directions.

Warm and Cold Currents

Currents may be classified as follows:

(*a*) Those of which the temperature corresponds to that of the latitude in which they flow; this temperature may be warm, cold or intermediate. Examples of these are the warm west-going Equatorial Currents and the cold east-going Southern Ocean Current, associated with an east–west trend of the sea surface isotherms.

(*b*) Those of which the temperature does not correspond to that of the latitude in which they flow. They are either warmer or colder than currents of class (*a*) in the same latitudes. Examples of these are the warm Gulf Stream and Kuro Shio, which bring water from the Equatorial Currents to higher latitudes, and the cold East Greenland Current, flowing from the Arctic Basin to lower latitudes. Owing to the transport in latitude of warm or cold water by these currents, the sea-surface isotherms trend northward or southward (*see* Chapter 8).

The cold currents of class (*b*) may be divided into two kinds:

(i) Those coming from polar regions to lower latitudes, such as the East Greenland Current, the Labrador Current, the Falkland Current and the Oya Shio. These do not form part of the main circulations round the permanent anticyclones.

(ii) Currents of lower latitudes, such as the Peru Current, forming the eastern part of the main circulation. In these cases the coldness is

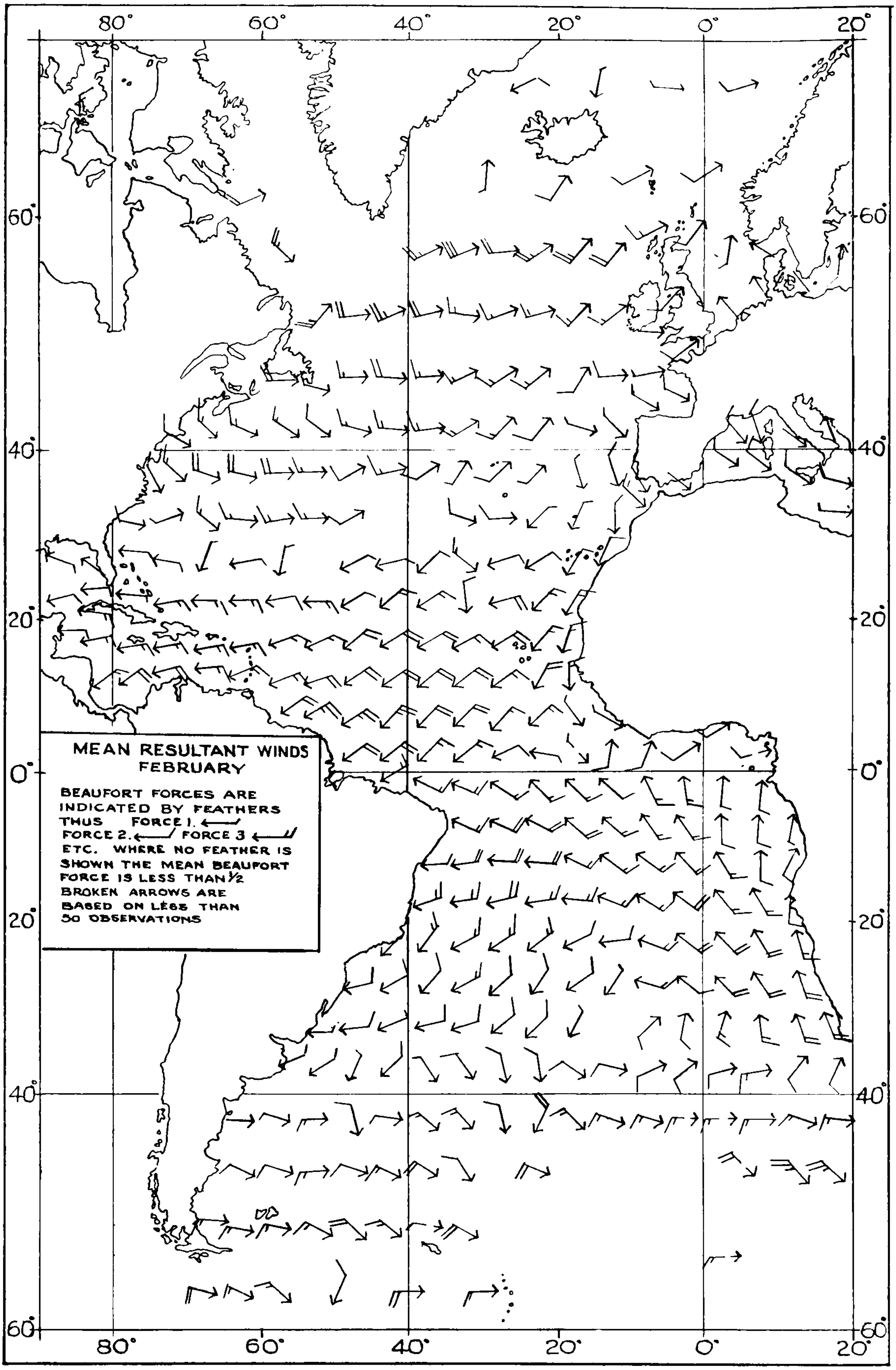


Fig. 16.1. Sea-surface winds over the Atlantic Ocean

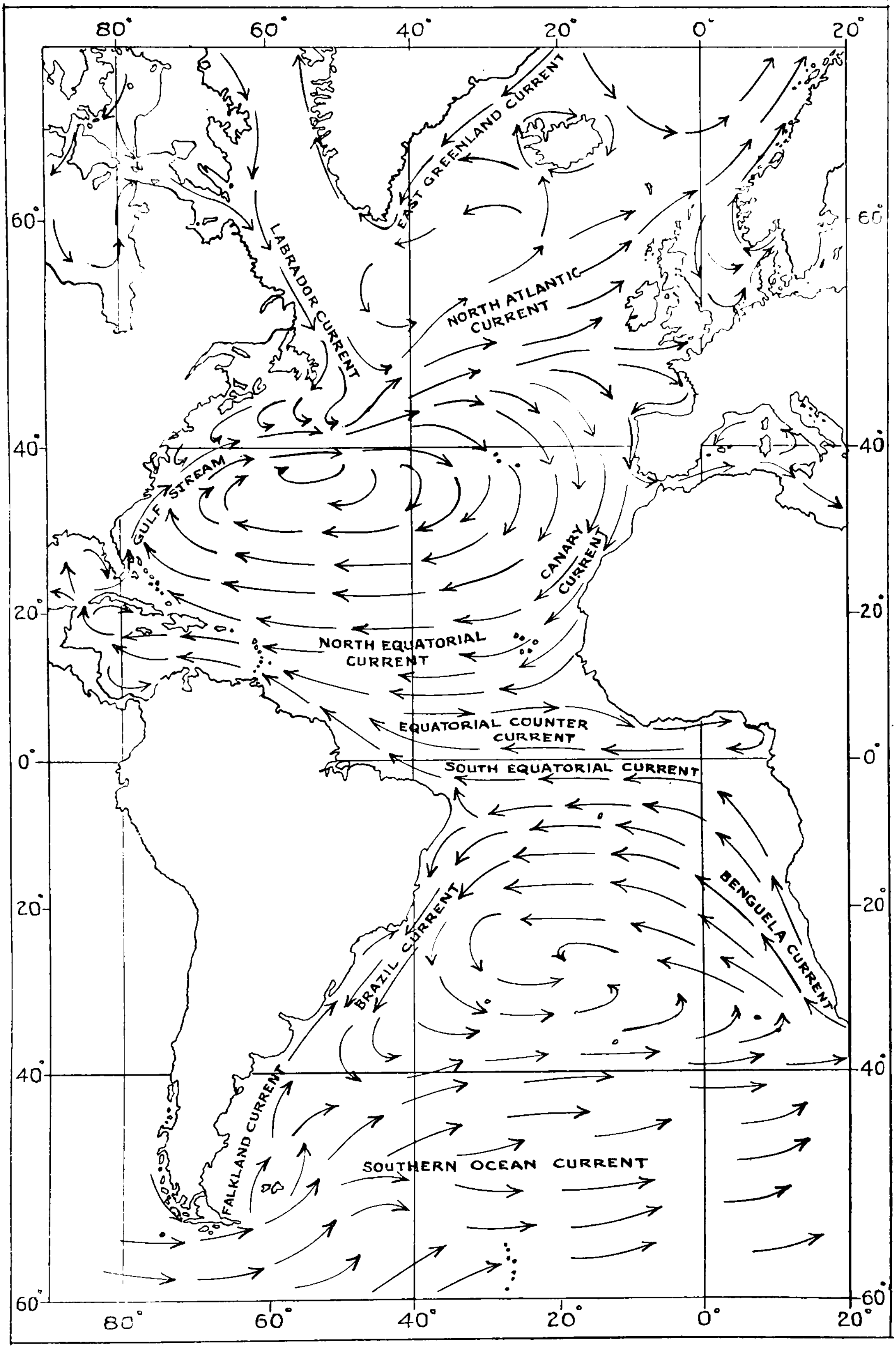
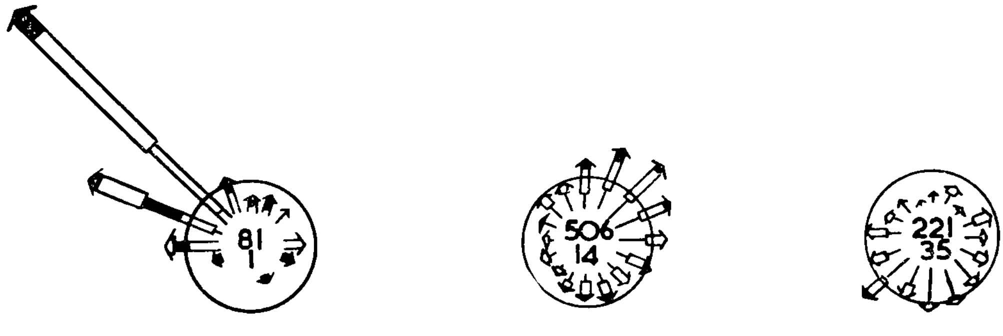


Fig. 16.2. Sea-surface currents for the Atlantic Ocean

mainly caused by water rising from subsurface depths, near an extended coastline. The 'upwelling' water is not as cold as that of the currents described under (i).



(a) (b) (c)

Fig. 16.3. Three typical current roses

The warm and cold currents of the western and eastern sides of the main circulations are as shown in Table 16.1:

Table 16.1. Warm and cold currents of the western and eastern sides of the main circulations

Ocean	Warm Current on Western Side of Ocean	Cold Current and Area of Upwelling on Eastern Side of Ocean
North Atlantic South Atlantic	Gulf Stream Brazil Current	Canary Current Benguela Current
North Pacific South Pacific	Kuro Shio East Australian Coast Current	California Current Peru Current
South Indian	Mozambique and Agulhas Currents	—

There is no upwelling on the east side of the South Indian Ocean, where no extended coastline occurs, but the West Australian Current is usually a little cooler than the normal for its latitude, as part of its water comes from the Southern Ocean Current. The relative warmth of the currents on the west sides of the oceans, compared with other water in the same latitude, is greatest in winter and least in summer.

Cold currents from high latitudes are significant to navigators:

(*a*) because they bring ice to lower latitudes;

(*b*) because they contribute to a high frequency of fog in some regions (*see* page 97).

Strength of Currents

This information is taken from the current atlases, and refers to the open ocean, mainly between lat. 50°N and 50°S. It does not refer to tidal streams, nor to the resultants of currents and tidal streams in coastal waters. Information as to current strength in higher latitudes is scanty.

The proportion of currents of less than a quarter of a knot varies considerably in different parts of the oceans. In the middle of the oceanic circulations, where current is most variable, the weakness of the resultant flow is not caused by an unduly high proportion of very weak currents, but by the varying directions of the currents. There is probably nowhere in the open ocean where currents do not at times attain a rate of at least 1 knot.

The region where currents of between 2 and 3 knots have been experienced are given below, together with the parts of those regions where the currents may reach a strength of over 3 knots. The information refers to all months of the year, unless otherwise stated. Very few currents exceeding 2 knots have been recorded in other parts of the oceans.

Currents 2–3 knots	**Currents over 3 knots**
ATLANTIC OCEAN	ATLANTIC OCEAN
(*a*) South Equatorial Current, mainly west of long. 32°W, and its extension in the southern part of the Caribbean Sea.	(*a*) South Equatorial Current, between long. 34°W and 70°W, except in February to April.
(*b*) Gulf Stream, from Florida Strait to long. 40°W.	(*b*) Gulf Stream, Florida Strait to long. 58°W.
(*c*) Guinea Current (but not the Equatorial Countercurrent as a whole).	(*c*) Guinea Current, May to July.
(*d*) Mediterranean, Gibraltar to long. 2°E, May to January.	(*d*) Eastward of Gibraltar, August to October.
(*e*) Falkland Current and its extension, the Brazil Inshore Countercurrent, lat. 42°S to 28°S, May to July.	—
(*f*) Vicinity of Cape Town.	—
INDIAN OCEAN	INDIAN OCEAN
(*a*) Equatorial Current in the region of Madagascar, most frequent in May to October.	(*a*) Equatorial Current in the region of Madagascar, May to October.
(*b*) Equatorial Countercurrent.	(*b*) Equatorial Countercurrent, an occasional observation.
(*c*) Mozambique Current and its extension, the Agulhas Current.	(*c*) Mozambique and Agulhas Currents, more frequently in the latter.
(*d*) East African Coast Current* in both monsoons, i.e. whether it is flowing northward or southward along the coast, including the region south of Socotra.	(*d*) East African Coast Current* (most frequent in May to July). The region south of Socotra, May to November.
(*e*) The South-West Monsoon Current in the Arabian Sea and Bay of Bengal.	—
(*f*) The region immediately east and south of Ceylon, throughout the year.	(*f*) The region immediately east and south of Ceylon, June to December.
NORTH PACIFIC OCEAN	NORTH PACIFIC OCEAN
(*a*) North Equatorial Current, west of long. 152°E, December to May.	—
(*b*) Equatorial Countercurrent, west of 140°E, and eastward of Mindinao and in the Celebes Sea, where the North Equatorial Current is recurving southward into the Countercurrent.	(*b*) Eastward of Mindinao, June to August.
(*c*) Kuro Shio, from Luzon to about long. 150°E (160°E in March to May).	(*c*) Kuro Shio, from eastward of Tai Wan to about long. 150°E, except in September to November when confined to the region of the south coast of Japan.
(*d*) In the China Seas, in both monsoon periods.	(*d*) Off the coast of Annam, August to December and in February. Very occasionally elsewhere in the China Seas.
(*e*) In the region of the Gulf of Panama, to long. 84°W, in November to July.	—
SOUTH PACIFIC OCEAN	SOUTH PACIFIC OCEAN
(*a*) South Equatorial Current, mainly on the eastern side of the ocean.	(*a*) South Equatorial Current, lat. 0° to 4°N. long. 92° to 110°W, August to October,
(*b*) East Australian Coast Current.	(*b*) East Australian Coast Current, north of lat. 34°S, October to April.

* Often known as the Somali Current.

The strongest individual currents reported to the Meteorological Office, and the month in which each was observed, are given below:

Gulf Stream, February, 136 miles per day, (about $5\frac{3}{4}$ knots).

Kuro Shio, November, 133 miles per day, (about $5\frac{1}{2}$ knots).

East Australian Coast Current, April, 96 miles per day, (4 knots).

Agulhas Current, September, 121 miles per day, (5 knots).

East African Coast Current* near the coast, September, 118 miles per day, (about 5 knots).

East African Coast Current* region, south of Socotra: September, 168 miles per day (7 knots); August and September, 144 miles per day (6 knots).

The greatest known current strength in the world thus occurs south of Socotra, between lat. 9°N and 11°N, during the height of the SW monsoon.

Direct Effect of Wind in Producing Current

The relation between the directions of the surface currents and the prevailing winds is closest in the open ocean. The coastal currents, forming the west and east sides of the main circulation, are due to other causes which will be discussed later.

When wind blows over an ocean the lowest layer of air in the boundary between atmosphere and ocean causes shearing stresses in the water and generates eddies and turbulence. The effect is to produce a tangential pressure upon the sea surface, augmented by the direct pressure of the wind on the waves. The rate of drift communicated to the water varies directly with the wind speed and inversely as the square root of the sine of the latitude, being about 2% of the wind speed in high latitudes and 4% in low latitudes. Thus, in low latitudes, the speed of a wind-produced current in miles per day is about the same as the wind speed in knots.

A current so formed is called a drift current. The effect of wind in producing current is not limited to this direct action, as will be explained later. Considering its direct and indirect effects, wind has by far the greatest effect in producing surface currents.

The most constant drift currents are those caused by permanent and semi-permanent winds. Next in importance are those resulting from a predominant wind, such as the westerlies of temperate latitudes. Transitory winds cause current variability. Transitory wind drifts are sometimes strong, and their effect varies between a slight change of current direction and its complete reversal. When the wind direction is most variable, so is the current.

When water in an ocean starts to move, because of the wind blowing at the time or any other cause, a deflecting force, due to the earth's rotation, comes into play. This force is to the right in the northern and to the left in the southern hemisphere. At the surface of an open ocean, the angle of deflection is 45°† in all latitudes, except very near the equator. For example, in the northern hemisphere, in the north-east trades, the surface current will set 45° to the right of south-west, i.e. west. In water of less depth the deflection is less, and in very

* Often known as the Somali Current.

† The figure of 45° given here and subsequently, is that originally derived by Ekman. More recent work suggests that the actual deflection at the surface of a deep ocean may be only about 30°.

shallow water may only be a few degrees, especially if influenced by the shape of the land.

The effect of wind is gradual, and the full strength of the current is not reached until it has been blowing fairly steadily for 24 hours or more. After the wind has ceased or changed in direction, the current due to the original wind may persist for a time, gradually weakening, before a new wind begins to establish a new current direction.

The direct wind effect on the sea surface does not extend to any great depth, and the depth depends on latitude, being greatest in low latitudes. It also varies directly with wind strength. For lat. 20° to 60° and a range of wind strength from Beaufort 3 to 8, the depth of the layer in which movement due to direct wind influence occurs varies from about 20 to 130 fathoms.

As already stated, the angle between the surface current and the wind producing it is 45° in deep water. Below the surface, the angle between current and wind direction increases with depth while the rate of current decreases. The motion of the surface water initiates movement in a shallow layer below, but the rate of movement in this layer is less because of the internal friction of water, and its direction is deflected to the right of that of the surface layer by the earth's rotation, in the northern hemisphere. Similarly, a weaker movement is initiated in a still deeper layer, with further deflection to the right and so on. Near the bottom of the layer affected by wind, the water movement is directly opposite to the wind direction, with a rate of only 1/23 of that of the surface water; this decreases to zero at the bottom of the layer.

The total effect of the wind is to give a resultant direction of movement of the whole layer of water at right angles to that of the wind, towards the right in the northern hemisphere and the left in the southern. This also applies to the gradient currents described below. Whenever a layer of water tends to move from any cause, the resultant direction of movement of the whole layer will be at right angles to that of the original tendency.

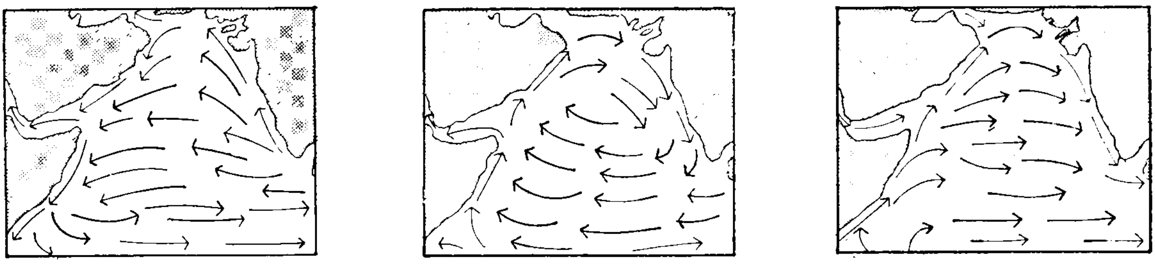
Gradient Currents

Currents known as gradient currents are produced, either at the surface or at any depth, when there is a pressure gradient in the water. At the surface, these are usually due to a slope of the water level, as occurs when water masses of differing densities lie adjacent to one another. Differences in temperature or salinity of the water cause density differences. A slope of the water may also be produced by wind.

The gradient current is produced as follows. Water tends to run down the slope, but immediately it starts to do so the effect of the earth's rotation diverts it to a direction at right angles to the slope.

A gradient current may flow at the surface at the same time as a drift current is being produced by wind, the surface current experienced being the resultant of the two.

An example of a thermal gradient current is found in the Arabian Sea and Bay of Bengal (*see* Fig. 16.4). During the NE monsoon, in December to January, the current in the open sea sets westward, with an anticlockwise circulation round the coasts. During February to April, the NE monsoon is still blowing, though it has weakened. A clockwise current circulation round the coasts becomes established during this time, while in the open sea it remains westerly. On the western coasts the current therefore sets against the wind. This coastal circulation is due to the cooling of the water at the head of the Arabian Sea and Bay of Bengal during November to January, while the NE monsoon is blowing offshore. The temperature difference sets up a slope, downhill towards the cooler water to the northward, and so produces the clockwise current. Towards the time of the onset of the SW monsoon the temperature difference decreases, but the monsoon arrives in time to preserve the current direction and to enhance its rate, as part of the normal SW monsoon circulation. Ships' observations clearly show the clockwise current in February to April.



(*a*) December to February (*b*) March and April (*c*) May to October

Fig. 16.4. Current circulation in the Arabian Sea

Effect of Wind Blowing over a Coastline

When a wind blows parallel to a coastline, or obliquely over it, a slope of the sea surface near the coast occurs and a gradient current results. A wind parallel to the coast is the most effective in creating a slope, since the total water movement, being 90° from the wind, is then directly on to, or away from, the coast. Whether the water is driven towards or away from the coast depends on (*a*) the direction of the wind along the coast, (*b*) the hemisphere in question. For example, in the Benguela Current (southern hemisphere) the SE trade wind blows obliquely to seaward over the coastline, i.e. towards north-west. The total transport of water is 90° to the left of this, i.e. towards south-west, and therefore water is driven away from the coast.

The coastal currents on the east sides of the oceans are produced in this way, by removal of water from the coastal regions by the trade winds. The gradient current runs at right angles to the slope, and as the slope is at right angles to the coastline the current is parallel to the coast. Taking the Benguela Current as an example, the water tending to run down the slope towards the coast is deviated 90° to the left and flows up the coast to the north (Fig. 16.5). At the same time the SE trade produces a surface drift current 45° to left of north-west, i.e. west; thus the actual Benguela Current experienced at the surface is the resultant of this and the gradient current, flowing approximately north-west.

Upwelling occurs in the coastal currents on the east sides of the oceans, since colder water rises from below to replace that drawn away from the coast by the wind. The balance between replacement of water by upwelling and its removal by wind is such that the slope of the surface and therefore the strength of the gradient current remains the same, so long as the wind direction and strength do not change. The slope is less than 1 inch in a distance of 10 miles.

The regions of upwelling on the east sides of the oceans and the resulting currents are as follows—the regions vary seasonally to some extent:

(*a*) The south-west coast of Africa, centred in about lat. 26°S. Benguela Current.

(*b*) The north-west coast of Africa, centred in about lat. 20°–22°N. Canary Current.

(*c*) The coasts of Chile and Peru. Peru Current.

(*d*) The coast of California. California Current. In this case there is no upwelling in winter months.

Two other regions where upwelling due to wind occurs on a smaller scale are the Arabian coast of the Gulf of Aden, during the SW monsoon, and the Gulf of Tehuantepec (Mexico) in winter, when 'northers' are blowing. Other types of upwelling are mentioned later.

The Labrador Current is produced by the banking of water against the coast by the northerly winds prevailing during much of the year. Water tends to run down the slope to seaward, and being diverted 90° to the right the current follows the coast southward.

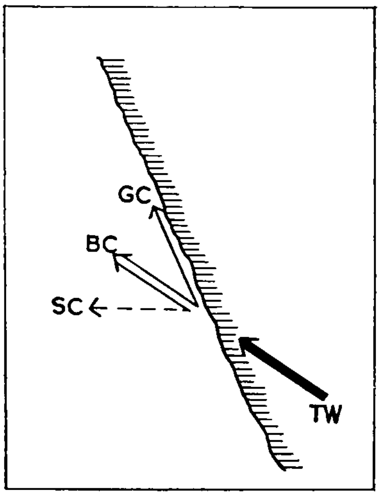


Fig. 16.5. Formation of the Benguela Current

TW is the direction of the trade wind blowing obliquely away from the coast, which lies in a NNW–SSE direction.
SC is the direction of the surface current that would be produced by the wind alone.
GC is the gradient current, surface and sub-surface, setting parallel to the coast.
BC is the actual direction of the Benguela Current at the surface, being the resultant of GC and SC.

Other Indirect Effects of Wind in Producing Current

We have seen that the direction of the surface current is 45° from that of the wind in the open ocean, also that the total transport of water by the wind, taking subsurface levels into account, is 90° from the wind direction. In the south-eastern part of the North Atlantic the north-east trade wind drives the surface water westward and the subsurface water north-westward. Thus water piles up in the centre of the main circulation eddy. This produces a gradient current which, tending to flow downhill to the south-east or south, is deflected 90° by the earth's rotation and so flows south-west or west, i.e. in nearly the same direction as the surface current directly produced by the wind. The drift current is thus strengthened by a gradient current, due indirectly to the wind.

The westward movement of water by the trade winds also produces a slope of the water right across the ocean in low latitudes. This uphill slope to the westward is probably less than 3 inches per 1,000 miles. Water thus accumulates on the western sides of the oceans. The Gulf of Mexico is a cul-de-sac in which water piles up to a measurable extent. It has been found that the sea level is

$7\frac{1}{2}$ inches higher at Cedar Keys, on the Gulf coast of Florida, than it is at St. Augustine, on the Atlantic coast of Florida. This head of water is the main cause of the flow of the Gulf Stream, starting through Florida Strait, which is the only possible exit. The other warm currents, on the west sides of the oceans are formed similarly by water accumulating against the coast. In some cases the wind blowing across the coast may enhance the current, as described earlier.

The Equatorial Countercurrents in each ocean flow eastward downhill in the doldrums. They are shallow in depth and are maintained by the piling up of the surface water against the west sides of the oceans.

Intermittent surface countercurrents are found between the main coastal currents on the west sides of the oceans and the coast. Thus, south of Cape Hatteras, southerly sets may occur between the Gulf Stream and the coast. Northerly sets are found sometimes between the Agulhas and Mozambique Currents and the coast. The Brazil Inshore Countercurrent, which flows north between the Brazil Current and the coast during part of the year, is described on page 205. These currents are cool and can be explained as part of the mechanism of the main current flow, whereby water rises from subsurface levels. On the other hand, the Caribbean Countercurrent is probably a surface eddy produced by the Equatorial Current in the Caribbean impinging against the indented east coast of Central America, part being thus diverted backward.

Mediterranean

In the Mediterranean the rate of evaporation is high, and the inflow of water from rivers is not sufficient to maintain the level of the sea. Water therefore flows in from the Atlantic, through the Strait of Gibraltar, to make good the deficiency. The effect of the earth's rotation is to tend to divert the east-going flow through the Strait to the south. There can be no actual deflection because of the narrowness of the Strait, but the inflow is thus kept against the whole length of the African coast and so gives a counterclockwise circulation. By evaporation, the Mediterranean surface water becomes saline and dense. It therefore sinks, and the excess of this denser bottom water emerges over the sill forming the shallow Strait of Gibraltar, below the incoming water. Being a subsurface flow this current, on entering the Atlantic, does not appear as part of the surface circulation. It is estimated that all the water in the Mediterranean is renewed in about 75 years.

Seasonal Variation of the General Circulation

(*a*) A complete reversal of direction occurs in different monsoon seasons.

(*b*) Seasonal variation occurs in most of the currents forming the general circulation. This may include change of direction, but is often mainly a change of average rate. Variations in the width or position of a current occur in some cases. The following are examples. The South Equatorial Currents in all oceans show a double maximum of strength during the year, though the months of greatest strength are not the same in different oceans. The boundary between the Guinea and South Equatorial Currents, on the Cape route, varies in latitude about 5° during the year. The origin of the Equatorial Countercurrent in the North Atlantic varies very much in longitude during the year.

(*c*) Certain coastal currents run for only a part of the year; these are referred to in the notes on the world current chart opposite page 202.

Variation of the General Circulation over a Period of Years

The current charts have been computed from all available observations extending over as many years as possible. Thus they do not show whether the general circulation varies from year to year or over periods of years. The number of current observations so far received each year is insufficient to construct charts for shorter periods. It is probable that such changes do occur and there is evidence that the Gulf Stream flow was stronger in 1910–31 than in 1932–39.

Subsurface Circulations

The subsurface circulations are imperfectly known, since the oceans are vast and oceanographical work is limited in time and place. The greatest variations of temperature and salinity occur in middle and lower latitudes, within a layer varying in depth from about 270 to 550 fathoms. This layer is known as the troposphere of the ocean (*see* page 287); it includes the still shallower stratum directly affected by wind. The strongest currents are confined to the troposphere. Below it the circulation in the open ocean is caused by lesser density differences and is relatively weak. The coastal currents on the west sides of the ocean, however, flow also in the deeper levels and perhaps nearly reach the bottom. These currents are deflected when they reach shallowing water, to the right in the northern and to the left in the southern hemispheres. Thus the Gulf Stream turns more to the eastward on meeting the Georges and Nova Scotian Banks, and the East African Coast Current* similarly turns seaward, southward of Socotra, on meeting the Carlsberg Ridge. (*See* page 261.)

The main surface circulation of an ocean, though it forms a closed eddy, is not self-compensating. The current charts show that the same volume of water is not being transported in all parts of the eddy; strong and weak parts can be found in all circulations. Also there is some interchange between different oceans at the surface. Thus a large part of the South Equatorial Current enters the North Atlantic to join the North Equatorial Current, and so enhances the Gulf Stream. Surface currents alone cannot adequately compensate for this; in addition to horizontal subsurface flows, there must therefore be interchange between surface and subsurface water. Upwelling near coasts has already been described; it also occurs elsewhere, and so does the opposite process of sinking. Water masses in the ocean, like air masses, can converge together or diverge from one another. At a convergence, water sinks below the surface to remove the excess; at a divergence, water rises from subsurface levels to compensate for that which is being removed laterally.

The Antarctic Convergence, where 'Antarctic' surface water sinks below the less dense subantarctic surface water lying northward of it, can be traced all round that continent, in varying latitudes, mainly between 50°S and 60°S. This is an interesting oceanographical feature but does not show in the current atlas as a convergence of surface current, for westerly winds prevail over the whole region, and the Southern Ocean Current flows eastward on both sides of the convergence.

A surface current convergence shown in the world current chart opposite page 202 is that between the cold Labrador Current and the warm Gulf Stream. This stretches from north of Cape Hatteras to Newfoundland. A similar convergence may also be seen between the cold Oya Shio and the warm Kuro

* Often known as the Somali Current.

Shio. The only other convergence appearing in the current atlases is the 'Subtropical' one in lat. 30°S to 35°S, in the eastern South Atlantic, and in lat. 35°S to 40°S in the extreme west of the South Indian Ocean. Here cool subantarctic surface water sinks below the warmer subtropical surface water to the northward.

Convergences and divergences occur in equatorial regions. In the Atlantic there is convergence at the boundary of the South Equatorial Current and the Equatorial Countercurrent in about lat. 4°N. At the equator and at the northern boundary of the Countercurrent, divergence takes place and water rises.

In all the above cases the rising water does not come from the bottom of the ocean, nor does the sinking water go to the bottom. The interchange is between the surface and moderate subsurface levels. In restricted areas in high latitudes water sinks the whole way from the surface to the bottom.

Oceanographers have found that there is a four-fold stratification of the waters of the oceans. Below the troposphere there is a layer of 'intermediate water'. Below this, in succession, are the 'deep water' and 'bottom water' layers. A description of the main subsurface flows in the Atlantic is given on page 289.

PART VI. ICE

CHAPTER 17

THE FORMATION AND CHARACTER OF SEA-ICE AND ICEBERGS

Classification of Ice

Ice formations may be classified as : (1) Ice of land origin. (2) River and lake ice. (3) Sea-ice.

Ice of land origin is divisible into:

(*a*) THE ICE CAP (ICE SHEET or CONTINENTAL ICE), such as covers the interior of Greenland and the Antarctic continent.

(*b*) GLACIER-ICE. This may extend out to sea in a mass of ice called an ICE-SHELF which, being continually pushed out from the land, floats as soon as deep enough water is reached. In some cases this floating ice extends seaward for a great distance compared with its width, being then known as a GLACIER TONGUE. Ice-shelves and glacier tongues of considerable size are found only in the Antarctic. ICEBERGS are masses of floating land ice derived from glaciers, where these reach the sea, or also, in the Antarctic, from ice-shelves or glacier tongues.

All ice originally formed in sea water is called SEA-ICE. It is classified as: (*a*) Sea-ice attached to the shore (FAST-ICE); (*b*) sea-ice which drifts freely under the action of current, wind or tidal streams.

Some of the Antarctic ice-shelves are formed by the annual accumulation of snow upon persistent sea-ice. An ice-shelf may, therefore, be either of glacier or sea-ice origin.

Of all the ice met in the open sea, the greater amount is sea-ice. Near a coast, may be met, and also in the Antarctic, land ice extending into the sea in the form of ice-shelves and glacier tongues. RIVER-ICE and LAKE ICE may occur in the northern hemisphere. The former originates as fresh-water ice and is usually in a state of decay by the time it reaches the open sea, so is only of local importance.

Ice Nomenclature

Ice takes many varied forms, and age and thickness are taken into account in naming these, also the relative amounts of ice-covered sea and open water, in relation to navigation. Ice nomenclature terms are therefore numerous and some of them have in the past been differently applied in various countries. The nomenclature used in some *Admiralty Pilots* and hitherto in general use by British seamen was given in Chapter 7 of the *Marine Observer's Handbook*, Seventh Edition. A committee of ice experts serving on the Commission for Maritime Meteorology have drawn up an International Ice Nomenclature and this was officially adopted by the World Meteorological Organization for international use in December 1955. This nomenclature is given in full in the Eighth Edition of the *Marine Observer's Handbook*; it does not differ essentially from that previously in use by British seamen. A few terms have been altered and in some cases the use of alternative names for the same ice form has been authorised. The number of terms has been somewhat increased and where possible the definitions have been made more precise. A shorter list of ice terms together with ice photographs is given in *The Mariners Handbook* (S.D. 100) published by the Hydrographic Department of the Navy.

Formation, Development and Decay of Sea-Ice

The freezing-point of sea water is lower than that of fresh water; the higher the salinity the lower the freezing-point. Average sea water, with a salinity of 35°/₀₀,* freezes in the open sea at a temperature of 28·6°F. The comparatively fresh water of the greater part of the Baltic, with a salinity of 5°/₀₀, freezes at about 31·5°F. The relation between salinity and freezing-point, and the effect of the temperature of maximum density of sea water on the rate of freezing, are described in Chapter 19, page 249. The deeper the sea, with waters of salinity less than 24·7°/₀₀, the later the time of freezing, or it may never freeze completely, as in the central part of the White Sea.

Ice forms first in shallow water near the coast or over shoals and banks, particularly in bays, inlets and straits where there is no current and where salinity is low, such as near river mouths. It spreads, and if broken up and carried seaward by wind or currents starts further ice formation in deeper water. Ice in deeper water not melted during the previous season also acts in the same way. Wave action ordinarily hinders the formation by mixing up the water of the upper layers. The presence of old ice damps down any sea or swell, and at the same time, by cooling the water, tends to assist the beginning of freezing. Recurring fresh winds hinder ice formation by breaking it up.

The first sign of freezing is an oily or opaque appearance of the water due to the formation of ice spicules and thin pointed plates about one-third of an inch across, known as ICE CRYSTALS or FRAZIL CRYSTALS. These consist of fresh ice free from salt, and increase in number until the sea is covered by SLUSH or SLUDGE of a thick, soupy consistency. Except in sheltered waters an even sheet of ice seldom forms immediately; the slush, as it thickens, breaks up, frequently into the characteristic PANCAKE form, the rounded shape and raised rim of which is due to the fragments colliding with each other. The formation of slush damps down any swell, and if the low temperature continues the pancakes adhere to each other, forming a continuous sheet. In pools and lanes of still water the sea freezes direct to BARE ICE. The final result in either case is YOUNG ICE, made up of ice plates about the size of a fingernail, set horizontally in the uppermost ½ in., but otherwise arranged in vertical bundles. Directly formed bare ice, 3 to 6 in. thick, cracks easily by reason of the plates being set vertically, and a ship can make better headway through it than through the slush and pancake ice of the open sea.

While the original ice plates or spicules are free from salt, a network of ice crystals forms as further plates are added, in the pockets of which sea water is trapped. Rapidly formed sea-ice holds more brine between its crystals than ice of slow growth. Young ice formed at an air temperature of 14°F has a salinity of 4 to 6°/₀₀, but that formed at −40°F may have salinity as high as 10 to 15°/₀₀. Ice is a poor conductor of heat, and its rate of formation drops appreciably after the first 4 to 6 in. have formed; a snow cover, if present, still further reduces the

* Parts per thousand (*see* page 250).

conductivity. The upper layers are therefore richer in salt than the more slowly formed layers. The enclosed brine is seldom frozen; it tends to sink through the surrounding crystal network because its density is greater than that of the ice around it. At low temperatures this draining process is slow, and in the first winter young ice loses only a portion of its salt. With the summer rise of temperature the process is more rapid, and LEVEL ICE, as young ice of this age is called, may lose almost all its salt. Young ice, while still moderately salt, is soft and damp to the touch. Horizontal bands may be conspicuous, marking varying rates of growth, but vertical streaking is probably the most characteristic structure in ice of this stage.

Sea-ice may grow to a thickness of 3 to 4 in. in the first 24 hours, and 2 to 3 in. more in the second 24 hours. With the subsequent falling rate of growth, ice which has grown steadily all the winter (WINTER-ICE) is seldom more than 4 to 5 ft thick by the following summer. The original ice surface is often below water level, owing to the weight of accumulated snow, and is rarely visible after the first few days.

In the Antarctic the snow cover remains unmelted during the summer following the original ice formation, and growth continues. Surface diatoms work into the ice in summer and are sealed in it in autumn, the resulting yellow layer showing that the ice is of more than one season's growth.

In the Arctic much higher summer temperatures prevail, and the snow cover melts and runs off or forms fresh-water pools. The sea round the floes may thus become covered with a thin layer of fresh water which spreads under the ice and freezes on the under side, in temperatures higher than those required to freeze salt water. In summer, therefore, an Arctic floe melts on top, but at the same time may slowly grow underneath. By this process mud, stones, seaweed or shells originally frozen to the under side of grounded floes may work right up to the surface. Diatoms frozen to the under side should similarly rise. An autumn period follows, with lower temperature but no ice formation, the supply of fresh water being no longer renewed and the sea temperature not being low enough for salt water freezing to begin. In the second winter growth continues by salt water freezing.

In the Arctic, ice may remain unbroken during its first summer. If it remains level and undisturbed for two winters, it will seldom be more than 7 to 8 ft thick. More usually the ice cracks and breaks up and is subjected to pressure, with resultant hummocking, either during the first winter or the following summer. Sea-ice, other than fast-ice in sheltered bays or along the coast, is continually in motion as a result of wind, tidal stream and current.

There are several factors tending to produce different motion in adjacent floes. Owing to differences of area and thickness, and the effect of wind and current on different masses of ice, floes travel at varying speeds. Wind and current are subject to local variations. In its motion the ice opens and shuts like a concertina; there are always leads and lanes present, otherwise it could not move. In summer the leads remain open except in very high latitudes, but in winter they are soon frozen over with young ice. Swell tends to break up ice, and in narrow or shallow waters tidal rise and fall will have the same effect. As a result ice is constantly breaking up, even in winter, and is frequently subjected to pressure. The onset or release of pressure may happen at any time, even in mid-winter.

As moving floes are driven together, or pressed against fast-ice, rafting or hummocking occurs, according to the degree of pressure. Ridges at right-angles to the direction of impact, or confused pressure areas of HUMMOCKED ICE, may be formed. The longer the pressure lasts the greater the chaos. Pressure ridges may be as high as 50 ft where grounded, but in deep water they are usually only about 10 to 15 ft high, sometimes reaching 20 to 30 ft. A ridge is at its highest when first formed, and some settlement soon takes place, owing to the weight of the hummocks. The weight of a ridge is ultimately supported by a downward extension of ice under the water which may be as much as four to five times the height of the ridge above. During summer, pressure ridges change in outline and the sharper features are softened to the form of rolling hillocks. Snowdrifts form against the ridges, the balance of weight alters, cracks form and the opening and closing process goes on.

The releases of pressure cause lines of weakness in icefields in the form of cracks, lanes or leads. These are often parallel to pressure ridges, but an icefield does not necessarily crack in its thinnest part.

Wind will tend to regroup ice that is scattered. As the wind rises the separate floes form lines in a direction at right-angles to that of the wind. These lines break up when the wind changes, and realign themselves at right angles to the new direction. When the wind blows offshore a channel of open water may be found between the coast and the ice, or may increase in width if already existing. A wind blowing on to a coast or on to fast-ice tends to reduce the width of any existing channel. If the wind is strong enough, hummocks may be produced along a line at right angles to the wind direction.

The air temperature as influenced by the wind also affects the grouping of ice. If the wind is a cold one, the lowered temperature may cause further freezing, so that the groups of ice may join together by new formation. In this case the ice would not be so readily broken up by a change of wind.

In the Antarctic it is unusual for sea-ice to be more than one to two years old. The drift in both the Weddell and Ross Seas carries the pack out into the open ocean in little over a year. In the Arctic, floes of much greater age are frequent. The heavy pack-ice of the Arctic Basin when of more than one winter's growth is called POLAR ICE. Ice formed off the Siberian coast takes about three to five years to drift across the polar basin and down the east coast of Greenland. Polar ice becomes pressed and hummocked to a degree unknown elsewhere but the warmth of the Arctic summers has its effect, and with polar ice more than two years old the hummocks are smoothed by melting to an undulating surface, which may be more or less level. Such ice is called ARCTIC PACK. The summer melting on the surface is considerable, about 2 ft as a rule, and pools of fresh water are formed on the floes. In the Antarctic, surface pools on floes in the pack are almost unknown.

The reduction in salt content of level winter-ice during the following summer was described earlier. If there is hummocking, this process is appreciably speeded up, and a single summer suffices for the ice to become fresh. This is accompanied by a change of texture and appearance, from greasy, streaky, opaque ice to a more or less granular, brittle and clear form. This ice is a source of the purest drinking water, having no taste of salt. Ice that remains unhummocked becomes fresh in its second summer. Fresh ice has a specific gravity of 0·916, that of young ice being about 0·925.

Wind has a direct influence on the melting of ice. Strong wind hastens the process by mechanical action, particularly if its direction is perpendicular to the ice-edge. In the Barents Sea, northerly winds in summer break off pieces of ice and drive them into the warmer water of the North Cape Current where they melt more rapidly. Southerly winds drive the warm water of this current up to the ice edge with the same result.

The final stages of melting vary with the type of ice. Fresh polar ice has a melting point near 32°F, whereas salty winter ice melts at lower temperatures. Ice of one winter's growth melts readily in low latitudes, if brine is still present. The internal melt due to variations in salt content produces a honeycombed appearance, and the ROTTEN ICE so formed soon disappears, the final stage being known as BRASH. Fresher and firmer ice lasts longer. Off the east coast of Greenland, polar floes alone reach and round Cape Farewell;* any young or winter-ice melts during the summer passage down the coast. In the last stages hummocked floes may persist as BERGY-BITS, about the size of a small cottage, and finally as GROWLERS, greenish in colour, awash and hardly visible.

Pack-ice

The term PACK-ICE (alternatively DRIFT ICE) is used in a wide sense to include any area of sea-ice, other than fast-ice, no matter what form it takes or how disposed. It is usually regarded by seamen, however, as a field of ice-floes which may be tightly packed or may be loosely disposed, with varying degrees of open water between the floes.

Pack-ice is classified as follows:

(*a*) VERY CLOSE PACK-ICE. Ice cover practically 8/8 and little if any water present.

(*b*) CLOSE PACK-ICE. Composed of floes mostly in contact. Ice cover 6/8 to 7/8.

(*c*) OPEN PACK-ICE. Floes seldom in contact with many leads and pools. Ice cover 3/8 to 5/8.

(*d*) VERY OPEN PACK-ICE. Water preponderates over ice. Ice cover 1/8 to 2/8.

The possibility of navigation through pack-ice is dependent not only on the closeness of the floes but also on their thickness and size. Unless the ice is light, i.e. thin, and the floes comparatively small, very close pack-ice is in general quite unnavigable except perhaps occasionally by powerful ice-breakers. With the same proviso, close pack-ice is usually only navigable by reinforced or specially constructed vessels and if the ice is very heavy, i.e. thick, icebreaker assistance may be required. Open pack-ice should be navigable with comparative ease by strengthened vessels and, if the ice is only 3/8 or 4/8, even by ordinary vessels, but caution and frequent changes of course are required. In very open pack-ice, if visibility is good, ordinary vessels can usually proceed at full speed, with few alterations of course. The polar ice of the Arctic basin is wholly unnavigable.

Fast-ice

FAST-ICE is sea-ice which remains fast in the position of growth, being attached to the shore along a coastline. The limits of fast-ice are governed by the configuration of the shore and the shelving of the bottom. Bays and inlets freeze up solid; islands lying near the shores are also usually within the fast-ice boundaries. Fast-ice moves up and down with the tides. The thaw starts on the shore and at river mouths and gradually spreads.

* These are called STORIS by the Danes.

Another form of fast-ice is known as the ICEFOOT. This is a low fringe of ice of varying height which skirts the shore, and being attached to the land is unmoved by tides and remains after the fast-ice has moved away.

Ice-shelves and Glacier Tongues of the Antarctic

The shelf ice pushes out seaward in large flat-topped masses, which float whereever the depth permits. This ice increases in height by the nanual accretion of snow on the surface. The fresh snow in time becomes coarse-grained and compact owing to temperature changes. In this state it is called NÉVÉ, the transition stage to glacier-ice. An ice-shelf may also be formed by snow accretion on seaice near the land.

The largest ice-shelf formation, called the Ross Ice-shelf,* was discovered in 1841. It stretches approximately east and west between Ross Island and King Edward VII Land, a distance of about 400 miles. Its greatest width extends about 500 miles from shore and its area is not far short of 150,000 square miles. Its average height above sea level is about 200 ft and it faces the sea as an ice cliff varying in height from 160 to 6 ft, which is almost everywhere afloat. The Ross Ice-shelf is one of the main sources of tabular bergs. Similar formations occur elsewhere in the Antarctic, but they are all much smaller than the Ross Ice-shelf.

A special feature of the Antarctic coast is the long, parallel-sided, floating tongues, extending seaward from glaciers and bounded by vertical 'ice-cliffs', the heights of which range from a few feet up to about 150 ft. Their lengths vary from a few miles upward. The Ninnis tongue, in Wilkes Land, extends northeastward 45 miles from Cape Spencer, and then for about the same distance north-westward; it is from 12 to 20 miles wide. The Drygalski tongue, the largest in the Ross Sea, is 38 miles long.

Icebergs

The Arctic bergs originate mainly from glaciers on the west and east coasts of Greenland, the country which contains 90% of the land ice of the north polar region. Of the bergs calved† by these glaciers, and reaching the open sea, nearly 70% come from the Disko Bay and North-east Bay region of West Greenland (lat. 69° to 71°N). The 12 chief glaciers of this area are estimated to discharge 5,400 bergs annually. The Devil's Thumb region in West Greenland (lat. 75°N) is also a very rich source of bergs. The most productive glacier on the more inaccessible East Greenland coast is believed to be that in Kangerdlugssuak Fjord in lat. 68°N. Many East Greenland bergs never reach the open sea; some are tightly sealed in by heavy pack-ice which may not break up for years at a time. In other cases icebergs are calved into fjords which they never leave, because of shallow water at the entrances, as happens with most of the big bergs in Scoresby Sound.

Svalbard (Spitsbergen) produces some small bergs, notably at the head of Stor Fjord and on the east coast of Nordaustland, and bergs are also calved in Franz Josef Land and the northern part of Nóvaya Zemlyá. Syévernaya Zemlyá, at the eastern extremity of the Kara Sea, probably produces more bergs than Spitsbergen or Franz Josef Land. Eastward of Cape Chelyûskin

* Formerly known as the Ross Barrier.
† A berg is said to be calved when it breaks away from its parent glacier.

(lat. 77° 43′N, long. 104° 16′E) not a single berg is produced along the north coast of Siberia, except those from small glaciers in the De Long Islands (a group of the New Siberian Islands). Bergs are calved in the inlets of the south-east coast of Alaska, but are rarely found far from the coast.

The breaking away of ice from the Antarctic continent takes place on a scale unknown elsewhere, and vast numbers of bergs are found in these waters.

The specific gravity of iceberg ice varies with the amount of imprisoned air. The mean value is estimated to be about 0·900, as compared with 0·916 for pure fresh-water ice. Many bergs show soft hues of green or blue; the more air it contains the whiter the berg. Owing to their large air-content tabular bergs of the Antarctic have a peculiar white lustre, as if formed of plaster of paris. Many glacier bergs appear dazzling white under certain conditions of light. The whiteness of a berg does not depend only on the amount of air originally in the ice; it is increased by weathering and by the effect of the sun, which releases innumerable air bubbles. Many glacier bergs show veins of soil or debris, while some have yellowish or brown stains.

The depth of a berg under water, compared with its height above water, varies with the type of berg and also with its age. Measurement of bergs off Newfoundland by the International Ice Patrol shows that about one-sixth of the berg may show above water in the case of the blocky, precipitous type of Arctic berg, and half in the later stages of a much-weathered berg. Growlers, which are more or less awash and waterlogged, have most of their mass below water and so constitute a danger to shipping.

Arctic bergs have varied shapes. Irregular dome-shaped bergs are the most numerous, but a flat-topped, precipitous-sided type also occurs, which is the nearest approach to the tabular bergs of the Antarctic. The height of Arctic bergs varies greatly: they frequently reach 230 ft and one of 447 ft has been measured. These figures refer to the height soon after calving; the decrease with lapse of time thereafter is usually rapid. The highest berg so far measured south of Newfoundland was 262 ft and the longest 1,696 ft. Bergs of over 3,000 ft long have been seen further north.

When ice is formed directly from snow lying on ice, as in the Antarctic, the crystals grow upwards from the ice beneath the snow. The air escapes as the crystals join up and the resulting ice, if very thick, appears blue. Bergs may be calved at any stage of this process, so they may be composed of frozen snow (névé) or hard glacier ice.

There are two types of Antarctic bergs:

(*a*) Tabular berg. This is the most common form, and it has no exact parallel in the Arctic. Flat-topped and rectangular in shape, hundreds of these bergs exceed a mile in length. Some are of great size, many up to 20 or 30 miles in length. The largest one reported was off the South Shetlands, in January 1927; it was about 100 miles square and from 100 to 130 ft high. The usual height of these bergs is between 40 and 120 ft, but greater heights have been measured, up to 200 ft. Névé tabular bergs consist of frozen snow, which gives them a white, woolly lustre. With weathering the colour may change to a soft blue.

(*b*) Glacier bergs, which are similar to those of the Arctic. Being of higher specific gravity than tabular bergs, they are more resistant to weathering. Ice island bergs are glacier bergs with conical summits and have been mistaken, especially when stranded, for islands covered with ice. Black-and-white bergs are a unique type of glacier berg, with distinct dark and white sections, observed only north of the Weddell Sea.

A berg is said to be weathered when it is in an advanced state of disintegration. Melting of the underwater surface is a continuous process and this, aided by mechanical action of the sea, produces caves or spurs near the waterline. This leads to a portion of the berg falling off or to a change in equilibrium, when tilting or capsizing may occur, thus presenting new surfaces to the action of sea and the weather. The presence of crevasses, earth particles or rock debris, facilitates melting or evaporation and produces lines of weakness, along which further breaking occurs. In grounding, a much crevassed berg may be wrecked or develop further cracks, which accelerate disintegration. The berg is finally disintegrated into relatively small pieces of ice known as growlers, which barely show above water.

For a complete description of signs of proximity of ice and of methods of navigation in ice, reference should be made to *Admiralty Pilots* for one of the Arctic regions or the Antarctic and *The Mariners Handbook* (S.D. 100).

CHAPTER 18

THE DISTRIBUTION AND MOVEMENT OF ICE

Introduction

Ice conditions vary considerably in different years. There are 'good ice-years' and 'bad ice-years', in which the extent of ice differs widely from the average. Abnormal years may come singly or as a period of successive years. A good or bad ice-year may be local, thus the bad ice-year, 1932, in Iceland was one of several good years in Svalbard (Spitsbergen).

There is evidence that the whole Arctic region has become slightly warmer in the last 30 years or so. This is shown by a decrease of the land glaciation of East Greenland and Svalbard and by some slight increase of mean sea-surface temperature. Whether this slow warming will continue is unknown, and it is possible that at any time a reversal may set in, with a gradual return to the previous conditions, or even to a colder state. During this 30-year period good and bad ice-years have continued to recur; for example, after a succession of good years in Svalbard the winter of 1940–41 was much more severe.

There is geological evidence that glaciation in Antarctica was, not very long ago, much greater than it is today.

The only way in which the distribution of Arctic ice can be shown is to base it on average conditions over a number of recent years. Figs. 18.1 and 18.2 show the average ice distribution in Arctic regions during the period 1920–40, for the month of greatest extent, April, and least extent, September. Figs. 18.4 and 18.5 show the average ice cover in the Antarctic since 1885, for the period August to September, when the pack ice extends furthest north, and February to March, when it lies furthest south.*

The 'extreme limits' of sea-ice and icebergs are those beyond which ice has only rarely been reported. Very occasionally ice is reported in abnormal places far beyond the lines shown as extreme limits. Thus in the North Atlantic it has been seen near the Azores, and as far south as lat. 28°N in the west of this ocean. In April 1894, an Antarctic iceberg was reported in lat. 26°S, long. 26°W, about the middle of the South Atlantic.

NORTHERN HEMISPHERE

Ice of the Arctic Basin

Polar ice (Arctic pack) covers about 2,000,000 square miles of the Arctic Ocean; its southern edge follows the 550-fathom line and it lies nearer to the Greenland-North America side than it does to Europe and Asia. North of Bering Strait and the north Alaskan coast, the pack extends to about lat. 72°N. Between Svalbard (Spitsbergen) and Franz Josef Land it does not reach the 80th parallel during an average year, although in exceptional years it may extend further south.

* Figs. 18.1 and 18.2 are taken from M.O.M. 390a, *Monthly Ice Charts of the Arctic Seas*. Figs. 18.4 and 18.5 are from Admiralty Chart 1241, *Ice Chart of the Southern Hemisphere*. The longer period is used for Antarctic ice because of the relative scarcity of observations in that area.

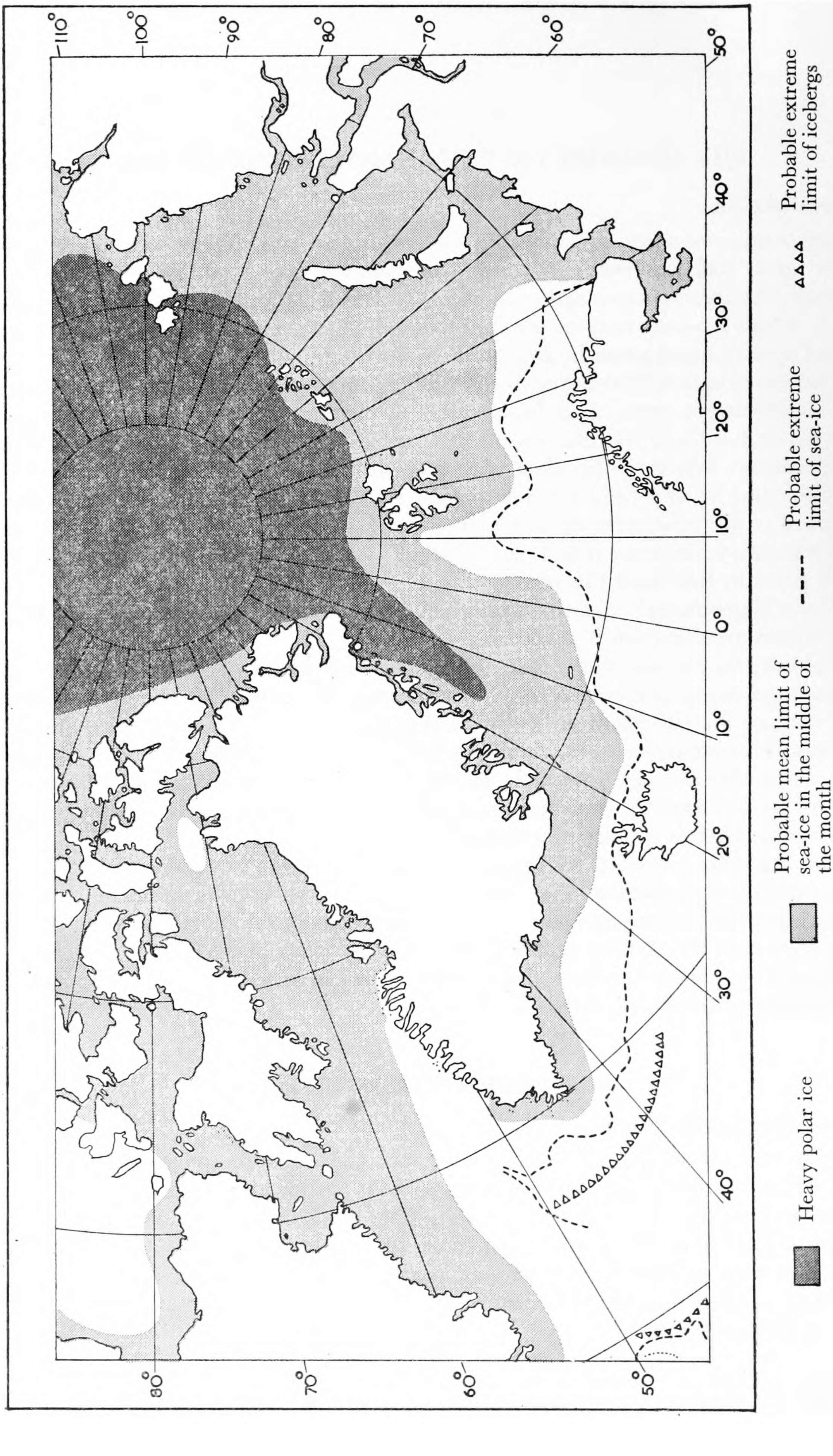
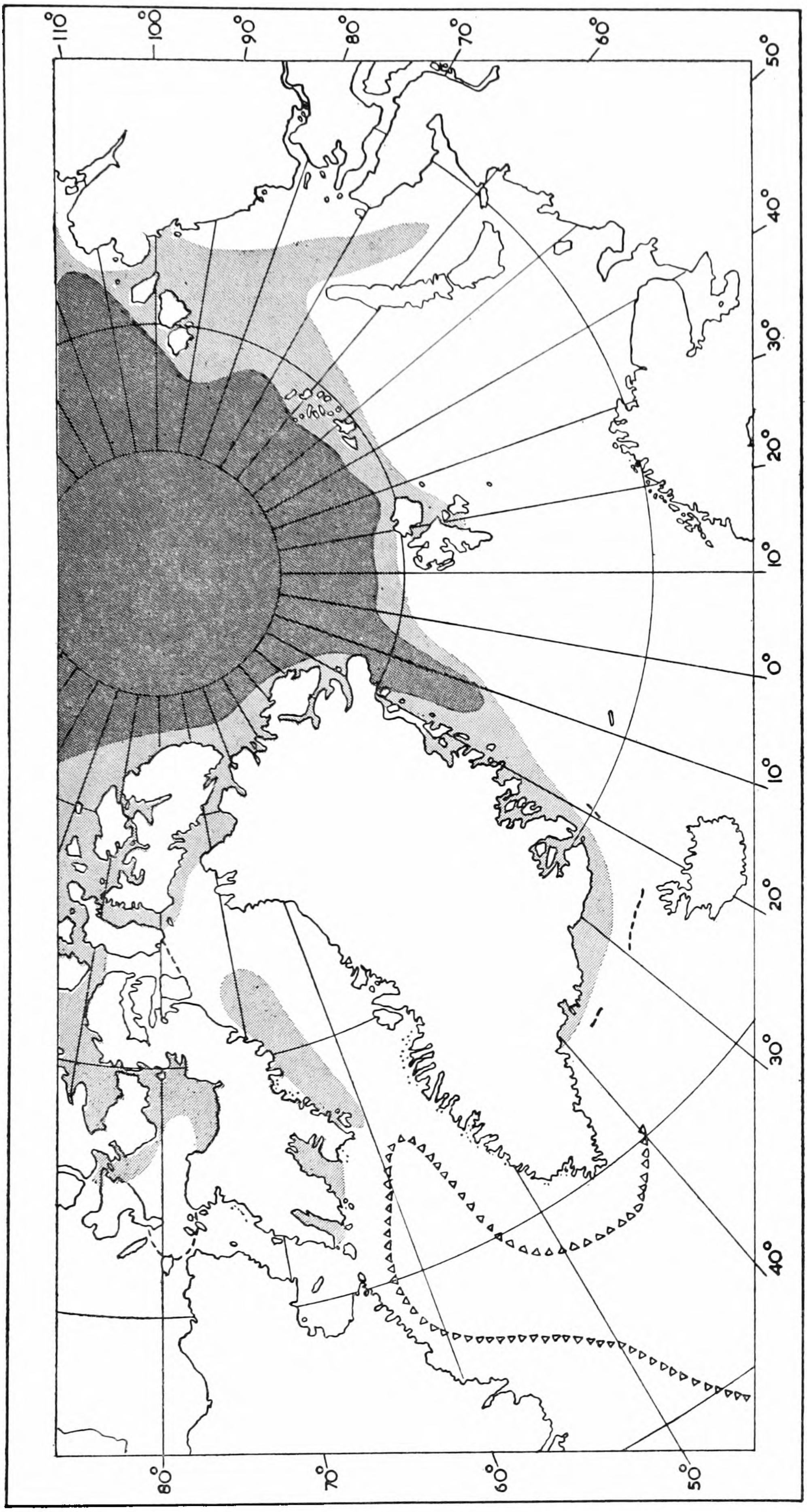


Fig. 18.1 Average conditions of ice in Arctic seas—April



Heavy polar ice

Probable mean limit of sea-ice in the middle of the month

Probable extreme limit of sea-ice

Probable extreme limit of icebergs

Fig. 18.2. Average conditions of ice in Arctic seas—September

Most of the polar ice consists of hummocked ice of many years' age. Sea water freezes in the Arctic Ocean to an average depth of 6 to 7 ft, but winds and currents produce hummocking and pressure ridges to a height of 30 to 40 ft and to depths of 100 to 200 ft. The ice cap is continually fed by pack-ice around its border and is also reinforced by new freezing and by snowfall. Freezing on the under side progresses during nine months of the year. On the other hand, ice is removed annually by melting and by discharge through the various exits into lower latitudes.

The currents of the Arctic Basin and those emerging from it are described on page 202. The general drift of ice within the Arctic Basin is therefore to the westward, as shown by the drifts of the *Jeanette* (1879–81), *Fram* (1893–95), *Karluk* (1913–14), *Maud* (1918–25) and *Sedov* (1937–40).

Emergence of Ice from the Arctic Basin

The main stream of ice from the Arctic Basin moves southward in the East Greenland Current. It is estimated that about one-third of the ice filling the Basin passes out of it in this way annually. As a result, a tongue-shaped mass of polar ice projects throughout the year in a southerly direction into the north-west part of the Greenland Sea. This tongue does not vary much in position during the year; it extends farthest south in April, in about lat. 74°N, and recedes farthest north in July to September, in about lat. 76°N. The longitude of the southern extremity of the tongue varies between about 11°W and 16°W during the year, being most easterly in August and September. The tongue touches the north-east coast of Greenland between north-east Foreland and about the 80th parallel, south of which it leaves the coast. Other ice, formed in the Greenland Sea, surrounds this tongue of polar ice, as shown in Figs. 18.1 and 18.2.

A second exit for ice from the Arctic Basin, on a much smaller scale, lies east of Svalbard (Spitsbergen), where the East Spitsbergen and Bear Island Currents carry ice south-westward into the northern part of the Barents Sea.

The easterly current flowing through the Canadian Archipelago brings some ice from the Arctic Basin in to the head of Baffin Bay.

Ice of the Greenland Sea

The western part of the Greenland Sea, owing to its lower sea temperature, has a much greater ice-cover than the eastern part. The cold East Greenland Current flows southward on the west side while the warm Norwegian Atlantic Current flows northward on the east side. Nowhere else can so high a latitude be reached throughout the year in ice-free water as off the west coast of Vestspitsbergen, where in mid-summer passenger vessels proceed north of the 80th parallel without much danger from ice. The west coast of Norway is ice-free all the winter.

Off Greenland, fast-ice may extend to about 10 miles from the coast and icebergs are found embedded in it. Outside the fast-ice is the pack-ice formed by the freezing of the sea *in situ*. Pack-ice thus fills the area between the fast-ice and the tongue of polar ice, off the north-east coast of Greenland. It also fills the rest of the northern part of the Greenland Sea, between the tongue and the open water off Vestspitsbergen, its extent decreasing to a minimum in September. Except near the ice-edge this pack-ice is unnavigable all the year. It may be up to 6½ ft thick and may reach a height of 25 to 30 ft in hummocks. The fields of unbroken close pack extend for distances up to 30 miles. Towards the ice-edge there may be considerable areas of more or less open pack, depending on the wind direction; the degree to which it is open varies in different years.

Pack-ice also extends down the east coast of Greenland, south of the polar tongue, the width of the belt varying seasonally. This ice, carried down by the East Greenland Current, is a mixture of polar ice floes, locally known as storis, broken away from the tongue and pack-ice formed locally. South of lat. 71°N breaks in the ice may occur in average years in August to October, and south of lat. 64° or 65°N there are large stretches of open water in September and October. Direct approach to the coast by ordinary vessels may be possible in August and September of average years as far north as lat. 71°N, while about lat. 75°N is the limit for specially protected vessels.

Movement of Arctic Icebergs

The general drift of bergs, once they are under way, is chiefly governed by the flow of ocean currents, but many factors produce different speeds of travel, such as wind, depth in water, local variations of currents, the stranding of some bergs, etc.

The main sources of origin of Greenland bergs, and their travel routes, are shown in Fig. 18.3. The bergs which reach the Newfoundland region and so become a danger to shipping originate from glaciers on the west coast of Greenland. The current carries the bergs northward from their birthplaces on the west coast of Greenland, round the head of Baffin Bay and then southward

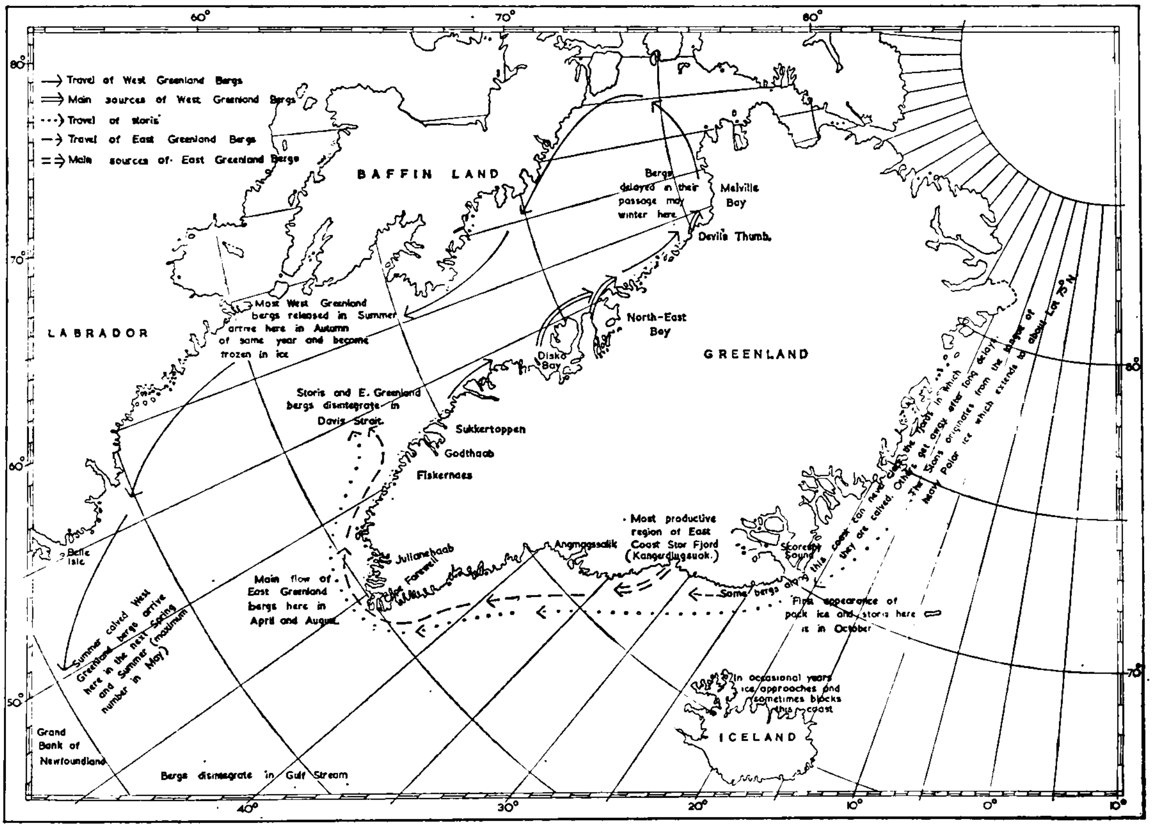


Fig. 18.3. Distribution and movement of Greenland icebergs

down the Labrador coast to the Newfoundland region. On arrival at the tail of the Great Bank, the bergs meet the warm water of the Gulf Stream, which hastens their disintegration.

The usual history of a berg is to be calved in summer, to reach the Hudson

Strait region the same autumn, to become fixed in the drifting pack and to appear off Newfoundland the following spring. A berg might, however, drift seaward during the summer instead of proceeding directly up the coast. Sooner or later it will re-enter the coastal current and may be held in the fast-ice of Melville Bay (west coast of Greenland). It is released from there in the second summer, reaches Cape Dier (Baffin Island) by October, and arrives south of Newfoundland with the main body of bergs in the next May, i.e. a total of about two years.

The first bergs reach the tail of the Great Bank early in April. The greatest number arrive in May. The average annual number observed south of the 48th parallel was 387 in the years 1913 to 1954 inclusive; May is usually the month of greatest frequency. Very great variations from this average number occur in individual years, ranging from over 1,000 to under 10. The years 1951 to 1953 were light ice-years, with 6, 14 and 56 bergs respectively coming south of the 48th parallel; in 1954 the number increased to 312.

The main flow of bergs from the east Greenland glaciers in the current round Cape Farewell is from April to August inclusive. Bergs may, however, be found in this locality in most months of the year. These bergs end by melting in the central part of Davis Strait, which is relatively warm at this season, and never reach the Great Bank of Newfoundland.

The bergs of Franz Josef Land and Svalbard do not usually travel far, but some may enter the East Greenland Current. Some Svalbard bergs drift south-westward in the Spitsbergen and Bear Island Currents and are usually found in small numbers near Bear Island from May to October. The Severnaya Zemlyá bergs are carried by the east-going current of the north Siberian coast down the east side of the Tamirski Peninsula into the Láptyevykh (Laptev) Sea.

Regions Wholly or Mainly Inaccessible at all Seasons

These comprise: (*a*) the Arctic Basin, (*b*) the north-east coast of Greenland, (*c*) the waters of the Canadian Archipelago.

Knowledge of the coast of Greenland north of about lat. 75°N has been obtained by passages in boats or native craft along the partly ice-free zone between the sea-ice and the land, or by means of sledge journeys on the ice.

The 'North-West Passage', through the Canadian Archipelago, has been made in both directions, but ice conditions do not admit of its being used for commercial purposes.

Ice in other Northern Regions

Davis Strait and Baffin Bay. These are never entirely ice-free in an average year; conditions are best in September. The east side of Davis Strait is free from about mid-May to November. The ice is a mixture of Arctic ice and that found locally.

Hudson Strait is generally open to navigation from the third week in July to mid-October. Recent work has shown that the whole of Hudson Bay is frozen solid from January to May, when it is covered with pack-ice; the entire bay is normally ice-free in August and September.

St. Lawrence River seldom, if ever, freezes over below Quebec, but during the winter and early spring is almost filled with ice that moves, with wind and tidal streams, from shore to shore. Above Quebec ice-jams occur; these caused destructive floods in former years but are now broken up by icebreakers. The average date on which the first vessel from sea arrives at Montreal is 28th April and that of the last departure for sea is 29th November.

Gulf of St. Lawrence is free of pack-ice from June to November in an average year, but in occasional years bergs may enter the Gulf from Belle Isle Strait during the whole of this period. Navigation is entirely suspended during the winter except for specially strengthened vessels or icebreakers.

Cabot Strait in most years cannot be navigated from the beginning of February until after the middle of April except by specially strengthened vessels. During part of April in most years, when the ice is clearing from the western part of the Gulf, Cabot Strait is wholly unnavigable, being filled by an ice-jam known as 'The Bridge', caused by a great rush of ice out of the Gulf.

Belle Isle Strait is normally open to navigation from mid-June to mid-December.

Nova Scotia. With the exception of Halifax, the harbours usually freeze during the coldest winter months. Navigation in Halifax harbour is scarcely ever interrupted.

United States, East Coast. Ice may form along this coast, especially in severe winters, as far south as parts of Chesapeake Bay and the estuaries opening thereinto. At the main ports steamer traffic keeps the channels open, with icebreaker assistance, when necessary, in the approaches to Philadelphia and Baltimore. Navigation is rarely interrupted and even in severe winters the interruptions last for only short periods.

White Sea is normally open to navigation from July to September, January to April being the worst months for ice.

The Northern Sea Route, from the White Sea to Bering Strait, is navigable during a limited period, with icebreaker assistance, but is never entirely ice-free. August and September are the best months, but there is considerable variation in different years, the navigation period varying from 70 to 120 days.

Baltic Sea. The severity of ice in the Baltic varies considerably in different years. The Gulfs of Bothnia, Finland and Riga are the regions most affected by ice; the worst month is March. The extreme southern part of the Baltic and the western part including the Kattegat are the least affected. The extreme north of the Gulf of Bothnia is ice-free only from July to October inclusive in average years. The main part of the Baltic, in normal years, has a narrow belt of ice along most of the coastal regions in the winter and early spring, the greater part of the sea being ice-free. Many of the Baltic ports are closed to navigation during much of the winter. Some are kept open by icebreaker when necessary.

Black Sea. Only the northern coasts are affected by ice, except in very severe winters. February is the worst month. The Sea of Azov is almost entirely ice-covered in February; the ice begins to form in November and does not disappear until towards the end of April.

Bering Sea is completely ice-free in normal years from about the beginning of August to early in October, except for the coastal region of Chukotski Peninsula. Maximum ice cover is in January and February, when the sea is mainly unnavigable north of the 60th parallel.

Sea of Okhotsk. There is some ice in coastal regions during all months except September. The worst period is January to April, but in the middle of the sea the ice is never heavy.

JAPAN SEA AND GULF OF TARTARY. The shores of the Gulf of Tartary are completely icebound from about mid-October to late in April. There is lighter ice and some open water in the centre of the gulf. South of the gulf some ice occurs on the western shore of Japan Sea. It is generally navigable except in the Vladivostok region, where a channel is kept open by icebreaker.

JAPAN. The northern and eastern coasts of Hokkaidō are mainly icebound in winter. Drifting pack-ice is found off the south coast of this island.

YELLOW SEA. Ice occurs round the northern coasts of this sea, north of about the 37th parallel. It is at a maximum in January and February, when the greatest amount of ice occurs in the Gulfs of Pohai and Liaotung. The entrance to Tientsin is kept open by icebreaker.

GREAT LAKES OF AMERICA. These are largely icebound from December to April, except for the southern part of Lake Michigan. Very powerful icebreakers are used on these lakes.

SOUTHERN HEMISPHERE

Distribution and Movement of Antarctic Ice

Conditions in the Antarctic differ widely from those in the Arctic. The circumpolar currents around the Antarctic continent explain the absence of heavy polar ice of many years growth.

In the Arctic Basin, the pack moves around a sea mainly surrounded by land, with one main exit. The Arctic pack may be travelled on without boats, and a relatively abundant animal life has enabled mariners shipwrecked there to reach civilisation. In the Antarctic, on the other hand, the drift of the pack is around, and outward from, an uninhabited continent towards open sea in every direction. These seas are the stormiest in the world. Landing is difficult on the continental coast, the greater part being bordered by high ice-cliffs. Even if a landing were made, the chance of rescue would be small, unless radio and speedy assistance from seaward were available. The only alternative would be to cross a wide belt of stormy and ice-encumbered sea to reach one of the few subantarctic islands.

The shores of Antarctica are surrounded by a medley of floating ice, comprising bergs and pack-ice, which forms an extensive ice-field, the forcing of which is attended with great difficulty and danger to any ship. The pack consists of sea-ice formed in the open sea, fast-ice formed along the coastline and also land ice.

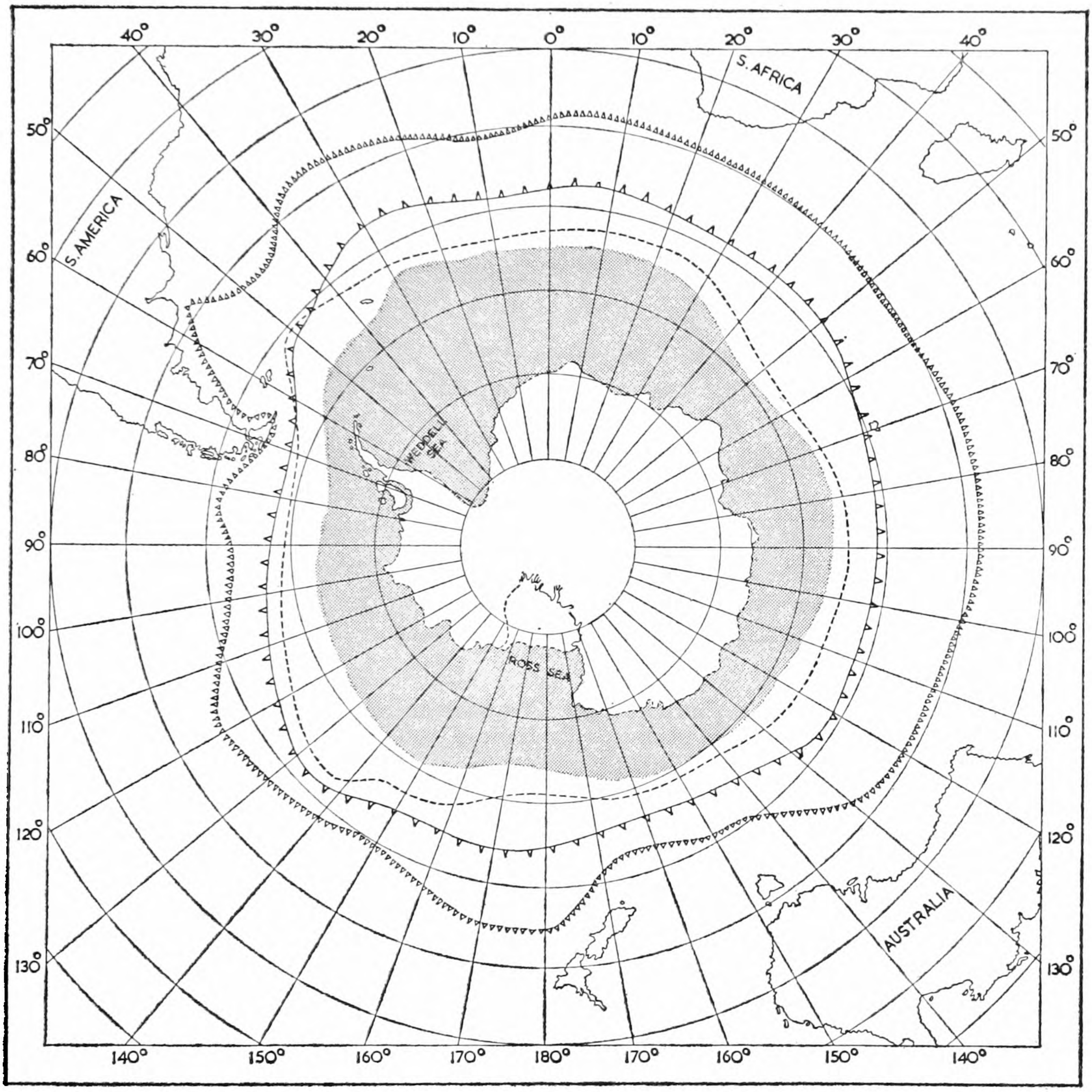
It contains ice of every size, from large floes to brash, with which bergs are mingled. The obstruction of capes and ice tongues in the paths of moving ice-fields, and the unequal movement between the floes, cause the ice to raft and hummock, and this, combined with the diverse character of the ice, prevents it freezing together and forming a solid mass during the winter. If this occurred, the ice might become permanent by the addition of snow and never be navigable.

The Antarctic sea-ice originates chiefly in the Weddell and Ross Seas, and along the coast of the continent. The freezing of the sea begins about the end of March. South-easterly winds assist the breaking up of the ice, and combined with the westerly current near the coast cause it to drift west and north. The pack from the Weddell Sea becomes heavily hummocked against the peninsula of Graham Land.

As the ice drifts out into the ocean new ice forms, which in turn follows the same drift. Throughout the winter the Antarctic seas are covered by moving pack which is continually breaking and becoming subject to pressure, due to wind, current, tide and drifting bergs.

The general drift of pack-ice near the continent is north of west. Further afield the north-westerly drift continues until, if not already melted, the ice reaches the region of westerly winds in about the 66th parallel. Here, owing to reversal of wind and current direction, the drift changes to about east-north-east.

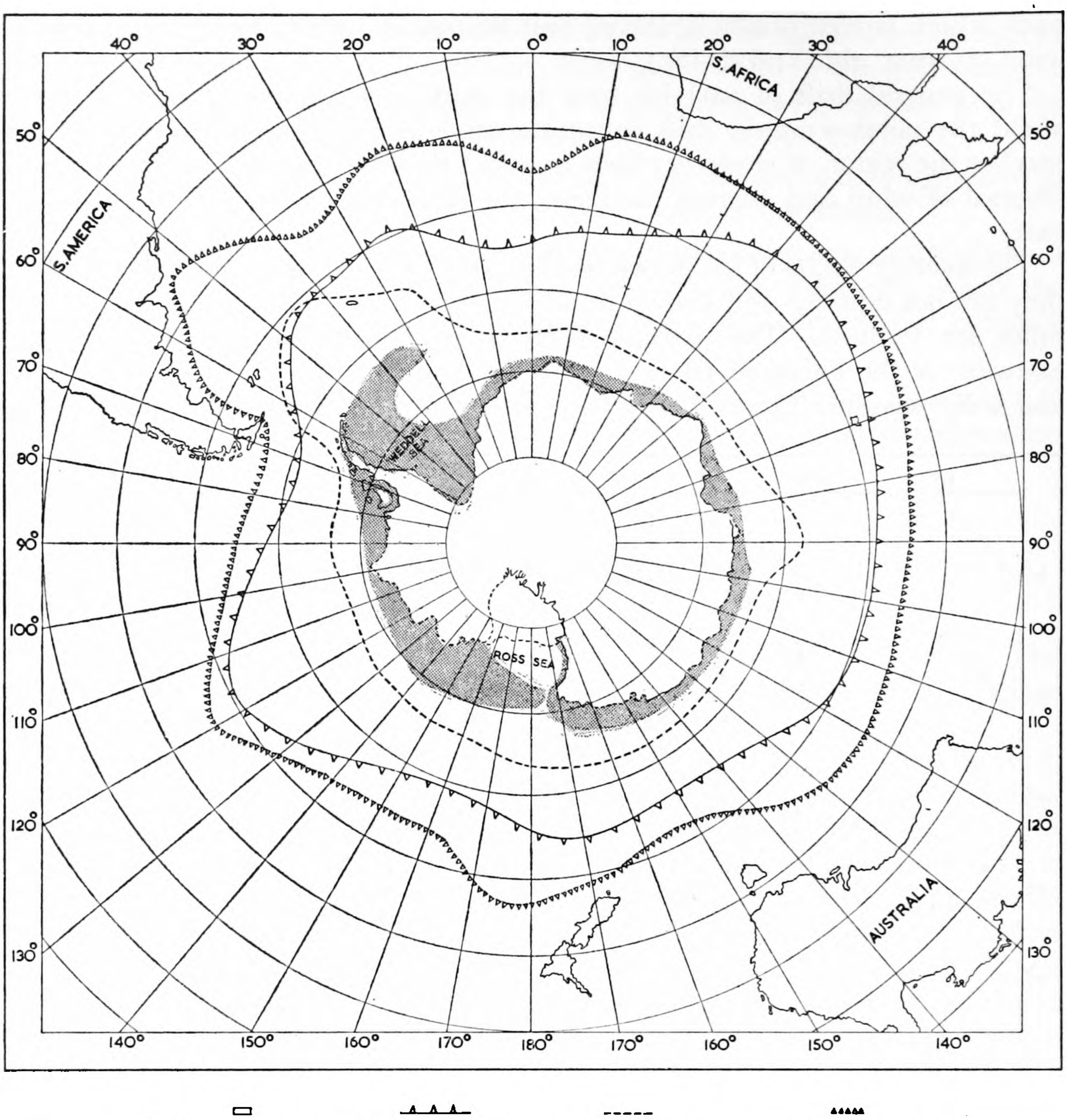
The average limits of ice shown in Figs. 18.4 and 18.5 are only approximate; they are not equally well known in all longitudes. Daily variations of 20 to 30 miles are frequent. The extreme limit of pack-ice is furthest north to the eastward of the Falkland Islands and is in about the same position in the best and worst seasons. This is due to the northerly trend of the Falkland Current.



Probable mean limit of pack-ice — Extreme limit of pack-ice — Probable mean limit of icebergs — Extreme limit of icebergs

Fig. 18.4. Average conditions of Antarctic ice—August–September

The period during which specially strengthened vessels have navigated the Antarctic pack is from January to March. The approach to the continent is most easily made in the Ross Sea region. Two landings in the southern part of



Probable mean limit of pack-ice | Extreme limit of pack-ice | Probable mean limit of icebergs | Extreme limit of icebergs

Fig. 18.5. Average conditions of Antarctic ice—February–March

the Weddell Sea were made in January 1956. The Royal Society Antarctic Expedition reached the coast at lat. 75° 36′s, long. 26° 45′w on the 11th and the British Trans-Antarctic Expedition landed at Vahsel Bay on the 28th.

Antarctic Icebergs

Owing to their depth under water, the movements of icebergs depend more on current than on wind. Generally they are almost unaffected by wind. Bergs in the pack will therefore often move in a different way to the pack itself. This relative movement helps to keep the pack open.

Because of their size icebergs normally last longer than pack-ice when air and sea temperature are increasing. The mean limit of icebergs extends much further north than that of pack-ice and is furthest north in November and December, about three months later than the greatest extent of pack-ice. It extends furthest north to about the 43rd parallel, in long. 45°w to 50°w and 10°E to 15°E. The mean limit is furthest south in May and June.

Two features of the Southern Ocean Current which affect the distribution of bergs are:

(*a*) After being constricted between South America and Graham Land it fans out north-eastward, extending furthest north in about lat. 38°S in the South Atlantic.

(*b*) The northward branch, known as the Falkland Current. As a result the extreme limits of bergs are further north in these regions than elsewhere; they occasionally reach the latitude of Rio de la Plata in December and January.

The frequency of bergs varies with longitude. They are most numerous in the western part of the South Atlantic, in the South Indian Ocean, between long. 40°E and 80°E, and in the South Pacific, between long. 90°W and 160°W.

PART VII. OUTLINE OF OCEANOGRAPHY*

INTRODUCTION

Oceanography as a Science

Oceanography emerged as a distinct branch of science in 1855, when Maury published his *Physical Geography of the Sea*. Modern oceanography dates from the expedition of H.M.S. *Challenger* in 1873–76. It is now a wide subject and only a brief outline of some of its aspects can be given here. The science deals with the physics and chemistry of sea water; it includes the study of the world below the sea surface and of the contact zone between sea and atmosphere. It thus embraces all the characteristics of the bottom and margins of the seas, and of the sea water with its living inhabitants. Oceanography is therefore associated with physics, chemistry, geology, the study of currents and marine zoology and botany. The relationship between oceanography and meteorology is also of importance, since the circulations of the oceans and the atmosphere are intimately connected, or have a large influence upon each other. The great mass of water in the hydrosphere, occupying a volume of about 330,000,000 cubic miles, thus directly or indirectly affects the life of every inhabitant of the earth.

Extent and Volume of the Oceans

The oceans and seas cover about 143,000,000 square miles, i.e. 71% of the earth's surface, and the average depth of the water envelope, the HYDROSPHERE, is about 12,500 ft (2,083 fm). If all the irregularities of the earth's surface were smoothed out there would be a water coverage over the whole of the resulting sphere some 7,500 ft (1,250 fm) in depth. The mean elevation of the land is about 2,300 ft above sea level. The greatest land elevation is that of Mount Everest, more than 29,000 ft. If we consider, however, the elevation of points on land above the bottom of the adjacent sea, Mauna Kea on Hawaii, which rises on that basis to a height of 31,750 ft, becomes the greatest mountain on earth.

The deepest part of the ocean is over a mile deeper than the height of the loftiest mountain. The existence in the Philippines Trench of depths in excess of 5,500 fm is well known. A strict evaluation of great ocean depths made in September 1950, accepted the Cape Johnson Deep in the Philippines Trench as the deepest at 5,740 fm, followed by the Emden Deep in the same general area at 5,687 fm. These two depths were taken by echo. The greatest accepted sounding made by wire up to that time was made by the Dutch vessel *Willebrord Snellius*, which plumbed 5,505 fm.

Since then, new information has come to light about very great depths in the oceans. The deepest spot now known lies in the same trench near the Marianas Islands in the Pacific Ocean. There a depth of 35,800 ft (5,980 fm, or about 6·8 miles) was reached by the bathyscaphe *Trieste* which took her two-man crew down to the trench bed and brought them safely up again. A series of corrections had to be made to the readings of the *Trieste's* pressure gauge to

* Excluding waves and currents which are dealt with in earlier chapters—the subject of tides is not dealt with in this book.

arrive at this depth figure and it has been preferred by one expert to call the depth 35·8 'kilo-feet' to convey that no significance attaches to the two noughts.

Little distant in time, American scientists made a sonic sounding of 35·7 'kilo-feet' from the research vessel *Stranger* in the same trench, and the Russians have reported a depth there greater by 370 feet than the *Trieste's* finding.

Area of the Oceans

Ocean	(A)	(B)	(C)
Pacific Ocean	63·8	69·4	35
Atlantic Ocean	31·8	41·1	21
Indian Ocean	28·3	28·9	15
Total	123·9	139·4	71

(A) The major areas of the oceans, in millions of square miles.

(B) The total area of the oceans (including connecting oceans and subsidiary seas) in millions of square miles.

(C) The total area of the oceans expressed as a percentage of the earth's surface.

CHAPTER 19

PHYSICAL AND CHEMICAL CHARACTERISTICS OF SEA WATER

The characteristics of sea water with which the oceanographer is mainly concerned are the temperature, degree of salinity and the relative density as compared with fresh water. He is also interested in the contents of the water, which include numerous salts, gases and substances such as phosphorus, iron, silica, magnesium, etc.

Temperature

In the surface layers of the open oceans the temperature varies between about 88°F near the equator, to about 29°F in the Arctic and Antarctic, this being the approximate freezing-point of the saltiest water found in polar regions. The distribution of sea surface temperature depends mainly upon latitude, the season of the year and ocean currents. Throughout the year the latitude effect generally dominates, as shown by the general temperature decrease from lower to higher latitudes. The average sea surface temperature is, however, higher in the northern hemisphere (about 66°F) than in the southern hemisphere (about 61°F). As a rough general rule, it may be taken that the surface temperature falls by 2/3°F for each degree increase in latitude, but in some localities the rate of fall may be much greater; south of Cape Horn it may exceed 2°F per degree of latitude. Furthermore, the rate of fall is slower in the tropics than in mid-latitudes.

If charts of surface currents and sea-surface isotherms are compared, it will be seen that the regional distribution of temperature depends greatly upon the 'run' of the currents, which in some areas, or in some seasons, determine the general trend of isotherms over considerable areas. In lower and middle latitudes there is an anticyclonic circulation of the waters in each ocean associated with a bending away from the equator of the sea surface isotherms in the west and a bending towards the equator in the east. The regions of pronounced upwelling, and resulting relatively cool surface water (*see* page 222), lie on the eastern sides of the oceans, and the bending of the isotherms in these regions causes them in some cases to run nearly north and south, as off the west coast of South America, and off south-west and north-west Africa and (in spring and summer) off the west coast of the United States. In higher latitudes, where the water circulation is dominantly cyclonic, this trend of the isotherms is reversed, the isotherms bending away from the equator in the eastern part of the ocean.

The temperature of surface water is generally highest in a belt just north of the equator; this shifts with the seasons but extends south of 'the Line' in only a few places. In most months, relatively low sea-surface temperatures are found along the equator, particularly in the Pacific. These are most marked where the Equatorial Countercurrent is best developed, and are associated with upwelling at a 'divergence' (*see* page 226).

On the western sides of the North Atlantic and North Pacific there are regions where sea surface temperature changes rapidly over short distances, due to the convergence of warm and cold currents. This is very prominent in the Atlantic, where the warm Gulf Stream meets the cold Labrador Current. In the Pacific it occurs with the warm Kuro Shio and cold Oya Shio. In the South Indian Ocean part of the warm Agulhas Current and the cold Southern Ocean Current come into proximity.

The annual range of monthly mean sea surface temperature is greatest in middle latitudes, particularly in the regions just referred to. Monthly mean air temperatures ashore can vary annually about 54°F in temperate regions, and more than 90°F in polar regions. Annual ranges of sea water temperature only amount to 9°F in the tropics and 18°F (at most) in temperate regions. Over large tropical areas, and in the Arctic and Antarctic, the annual range of monthly mean sea temperature is less than 4°F.

The extreme range of sea surface temperature is from 28·4°F to about 104°F; that of air temperature at surface levels (ashore) is about −90°F to 136°F.

Because of the high specific heat of water, the diurnal variation in sea surface temperature is very small. The average for all oceans is estimated to be 1·8°F. In the trade-wind regions it is only from 0·4° to 0·6°. In middle and higher latitudes the daily range is not constant through the year, but even at mid-summer it does not average more than from 0·6° to 0·9°, though with a clear sky on a calm day it might be as much as 4°. The water is warmest at about 1400 to 1500 local time and coolest at about 0500.

Salinity and the Chemical Composition of Sea Water

The term SALINITY is used to denote the content of all salts dissolved in sea water. Salinity is always expressed as parts per mille (=per thousand), for which the symbol ‰ is used. (*Note*—not %.)

In 100 parts of sea salt there are 77·8 parts of common salt (sodium chloride), 10·9 of magnesium chloride, 4·7 of magnesium sulphate, 3·6 of calcium sulphate (gypsum), 2·5 of potassium sulphate and 0·5 part made up of many other salts. Although the total amounts of the salts in sea water can differ greatly with the region (for instance, the Baltic and the Red Sea), the proportions of the various constituent salts remain the same, except for the concentrations of such substances as phosphates, which are affected by biological processes. If the content of one of the major constituents in a sea-water sample is determined, the content of any other can thus be calculated.

Nearly every known mineral element is present in sea water, some in commercially procurable quantities. The sea is now our main source of supply of metallic magnesium and bromine, the latter so important in connection with internal combustion-engine fuels. Sodium and potassium are being extracted but, like iodine, potassium cannot be extracted profitably except from certain seaweeds. Sea water itself contains only 0·037% of potassium, but dried kelp contains from 10 to 14%. Similarly, there is only 5 ten-millionths% of iodine in sea water, but in certain seaweeds (when dried) the content is from 0·2 to 0·5%; thus, at present-day prices, a cubic mile of sea water might contain iodine to the value of £15,000,000. It has been computed that the oceans contain about 120 thousand million tons of bromine compounds, and that 70 tons of bromine are present in 1,000,000 tons of water. Of the 22,000,000 tons of common salt estimated to be used by the world each year, only a small proportion comes direct from the sea, but a cubic mile of sea water (weighing thousands of millions of tons) contains approximately 117,000,000 tons of salt, 6,000,000 tons of magnesium, 283,000 tons of bromine, 192 tons of iodine, 550 tons of copper and also potassium salts in amount which could supply industry with 4,000,000 tons of lye. Finally, there would be about 3 tons of gold, which cannot be extracted except at a cost far greater than its value.

In the surface layers of the open oceans the salinity varies between 32 and 37·5‰, 35 being the average. Higher surface salinities occur in partly enclosed waters in low latitudes, where evaporation is excessive; the greatest occur in the Red Sea (up to 41‰) and the Persian Gulf. The salinity of the Dead Sea averages about eight times that of the open oceans, varying from about 192 to 260‰. On the other hand, in seas which receive much land drainage and are not subject to great evaporation, the salinity may be very small. Thus in the northern part of the Gulf of Bothnia the salinity may be no more than 3‰ and can be lower. In any latitude the water may be practically fresh at considerable distances off mouths of large rivers.

Of the oceans, the Atlantic has the highest average salinity, 35·37‰. Salinities are increased by factors which increase the rate of evaporation, such as those tending to warm the sea surface or move the air in contact with it. The higher salinities occur in the trade-wind belts, where warm winds are strong and there is bright sunshine. In each ocean there are thus two maxima of salinity, one in the northern tropical belt at about lat. 25°N and the other in the southern tropical belt, lat. 15° to 20°S. These are separated by an equatorial zone of lesser salinity, where there is much rain and generally little wind. The polar seas are regions of still lower salinity. In polar seas the surface water salinity will be appreciably less when ice is melting and greater when freezing is in progress.

Evaporation from salt water is only about 70 to 90% of that from fresh, in similar conditions. The total evaporation from sea water is estimated to equal the removal of a layer, about 40 in. thick, from all oceans in the course of a year. This gives an average, for the whole world, of about $\frac{1}{10}$ in. per day. The amount of evaporation varies greatly in different latitudes. Between lat. 10° and 20° the daily amount is estimated to be $1\frac{2}{3}$ in., falling off to as little as $\frac{1}{100}$ in., poleward of the 60th parallel. Annual evaporation in the Red Sea is very great and has been computed to equal a layer of water about 12 ft deep. Without replacement from the Indian Ocean the Red Sea would, in about 2,000 years, become a mass of salt.

The computation of current speeds by oceanographical methods often calls for very precise salinity data. Determinations of salinity are made chemically to determine the chlorine content when exactness is essential. This is convertible into salinity from the relation:

$$\text{Salinity} = 0{\cdot}03 + 1{\cdot}805 \times \text{chlorine content.}$$

The chlorine content refers not to chlorine alone, but also to the three other elements of the halogen family, bromine, fluorine and iodine, which are present in the sea salts. During recent years, however, there has been an increasing departure by professional oceanographers from the chemical method in favour of more precise and less arduous salinity determinations made via measurements of the electrical conductivity of samples of seawater. These are carried out by the use of seagoing electronic salinity meters and, to some extent, salinity can now be measured, *in situ*, by salinity probes. For precision work the concept of 'salinity' itself is also under review but the issues involved need not be considered here.

The rivers of the world are estimated to bring some 6,500 cubic miles of fresh water into the oceans each year, containing dissolved and suspended material to the amount of 2,700 to 5,000 million tons. The chemicals brought by the rivers are rich in carbonates and sulphates and poor in chlorides, whereas the chemicals in the sea are rich in chlorides and poor in carbonates and sulphates. Attempts have been made to estimate the age of the oceans by the time it would have taken for the rivers to supply all the chemicals at present in solution in the oceans. If our knowledge of the past distributions of land and water be taken into account, the present rate of annual addition of salts to the sea is not inconsistent with an oceanic age of a few thousands of millions of years. The age of the earth is considered on astronomical and other grounds to be of the order of 3,000 to 4,000 million years.

The distribution of nitrogen and phosphorus compounds in the oceans is very important; we shall return to this later, in its relation to life in the sea.

Density

The DENSITY of water or any other substance is its mass per unit volume. The density of a sample of sea water depends upon its temperature and salinity and, when it is below the surface, upon the amount of additional pressure to which it is subjected by the weight of the column of the water above it. Knowing these three quantities, the density can be read off conveniently from graphs, of which Fig. 19.1 is a specimen.

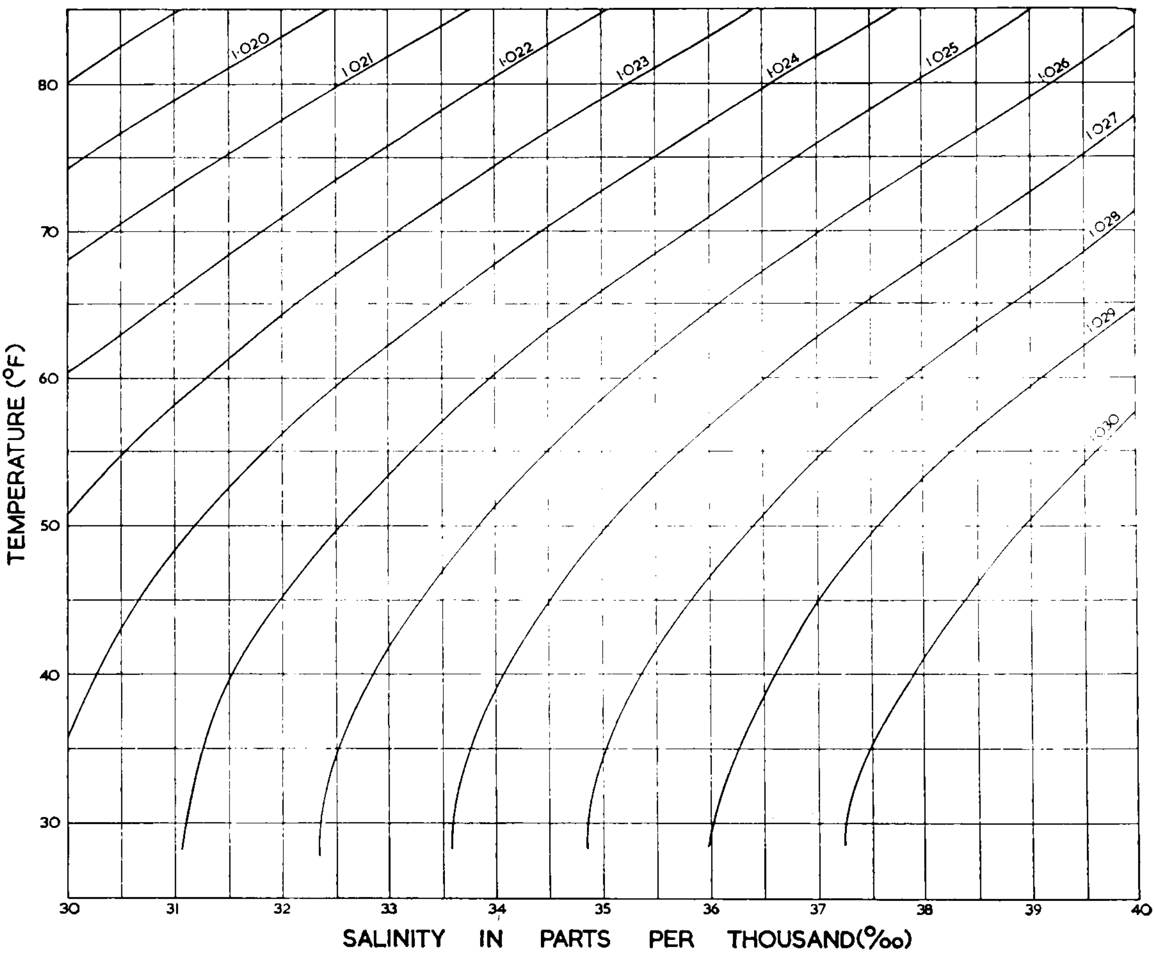


Fig. 19.1. Graph showing how to determine the density of sea water from its temperature and salinity

At a given temperature, sea water is heavier than fresh water because of the substances dissolved in it.

The specific gravity of sea water is a numerical ratio between the weight of a quantity of sea water and the weight of the same quantity of fresh water at the temperature of 39·2°F (4°C), at a pressure of one atmosphere (14·69 lb/sq in. or 1013·2 mb). The specific gravity of the surface waters of the open oceans varies between 1·022 and 1·028. Whereas a temperature change from 28° to 90°F will only alter the specific gravity by 0·007, a change of salinity from 0 to 40‰ would alter it by about 0·03.

At a temperature of 39·2°F and a pressure of one atmosphere the weight of 1 cu cm of pure fresh water is 1 gm, and the weight of the same volume of sea water in the same conditions in the open ocean is therefore from 1·022 to 1·028 gm. Thus specific gravity and density, expressed as weight/cu cm are numerically the same.

In oceanography, the density is expressed in a contracted form by using 1,000 times specific gravity reduced by unity; thus 1·022 would be written as 22·0. If the specific gravity of a sample of sea water is 1·02659, the value 26·59 would be referred to as the density, though in this form it is neither density nor specific gravity, but expresses how many more oz/cu ft the water sample weighs than does fresh water. Admiralty practice is to multiply the specific gravity by 1,000, thus: 1026·59, which indicates the weight of the sea-water sample as follows:

1026·59 oz/cu ft
1026·59 lb/100 gall
and 1026·59 kg/cu metre.

A ship moving from an area where the water has (in these units) a density of a certain value (say 1025·0) into another area where the density is one unit less (say 1024·0), will be 'tons heavy' to the extent of 1/1,000 of her displacement. If the density had increased by the same amount, she would have been 'tons light' to the same degree. Extremes of density difference may cause a difference in draught of some inches, thus there can easily be a difference of 8 in., as between the Baltic and the Red Sea.

The average density of the entire hydrosphere (kg/cu metre) has been computed as 1036, and the greatest water density *in situ*, i.e. at the greatest ocean depth, would be about 1077, allowing for compression of the water. Each increase in depth of 5½ fm adds one atmosphere of pressure. Pressure increases at the rate of one ton to the square inch for each thousand fathoms and is therefore 13,500 lb (6 tons) per square inch at the greatest depth. A ship sinking downwards might be squeezed flat if all its compartments were not filled with water. It is estimated that the level of the ocean surface would be about 98 ft higher than it is if water were entirely incompressible.

The old question as to whether objects could sink right to the bottom of the deepest ocean is still sometimes raised, but obviously the depths were only found in the first place by letting leads down to the bottom. Any object having a greater weight than the same volume of sea water has at a certain depth will sink beyond that depth, and anything heavier than 1077 oz/cu ft would go right to the bottom of the greatest ocean depth.

The use of hydrometers to measure density is familiar to seamen. Oceanographers also use hydrometers when great accuracy is not needed; for instance, in areas where conditions change rapidly and where bodies of water differing greatly in salinity are found near together. Though the common hydrometer is a simple instrument, it should be used with care, and as it is calibrated for a standard temperature, the temperature should be read at the same time, so that the recorded density can be corrected. Care should also be taken that no air bubbles attach to the instrument, and that the glass is clean and free from grease. Both these faults make the hydrometer more buoyant than it should be and the water appear denser than it really is.

The oceanographer favours a 'total-immersion' hydrometer, with which the entire instrument remains submerged, different lengths of very fine chain being lifted from the bottom of the vessel in proportion to water density.

Freezing-point

The freezing-point of sea water is lower than that of fresh water; the more so the higher the salinity. It is accurate enough to calculate the freezing temperature (°F) of sea water as 32°—0·095S, where S stands for salinity in parts per thousand. Water having a salinity of 35‰ starts to freeze at 28·6°F. Since salinity is never high enough (except in localised regions) to depress the freezing-point below 28·4°F, this is about the lowest temperature found anywhere in the sea.

Fresh water has its maximum density at 39·2°F. The temperature of maximum density of sea water is lower than this in proportion to the degree of salinity and can, with sufficient accuracy, be put at 39·2—0·387S (°F), where S is salinity. The temperatures of maximum density and of freezing-point coincide for sea water having a salinity of 24·7‰, which freezes at 29·6°F. The relation between the temperature of maximum density and that of freezing-point is important when considering ice formation in the sea, since the rate of freezing depends upon it. If the surface of fresh water cools to 39·2°F from any higher temperature, the density of the surface layer increases to its maximum and sinks through the lighter water below. More cooling will not further increase the density of the surface water, and as soon as the whole body of the water is cooled to 39·2°F further surface cooling produces no more convectional descent. Thus the temperature of the surface water begins to fall rapidly, and ice is formed as soon as this reaches 32°F. On the other hand the density of all sea water having a salinity of 24·7‰ or more increases with fall of temperature at all temperatures above the freezing-point of the water. Hence the convectional descent of the surface layers continues, and the rate of fall of surface temperature is delayed until all the water below the surface has reached its maximum density; ice therefore takes longer to form than in fresh water. In both fresh and salt water the surface freezes more rapidly if the depth is shallow; at the same temperature the surface of deep water may never freeze at all, since the duration of the cold period may not be long enough for all the water down to the bottom to reach the state of maximum density.

Most of the other physical properties of sea water can also be related to salinity (or chlorinity). There is a relation, for example, between salinity and the freezing and boiling-points of sea water, and when one of these is known the other can be calculated. Knowing the temperature of water *in situ* (i.e. in its original location at any depth) and its salinity, the density may be determined from Knudsen's *Hydrographic Tables*. Density figures are used for many purposes, for example to calculate the speed of currents.

Since water at any depth below the sea surface is compressed in proportion to the weight of the water column above it, a sample of water, on being raised to the surface, would expand and consequently become less dense. Though the water compression is slight at any depth, it is important, and therefore the oceanographer, after determining the temperature and salinity of a sample of water brought to the surface, applies a correction to the resulting density if he wishes to find the density the water had at its original depth. The uncorrected density of the sample at the surface is known as the POTENTIAL DENSITY, and this is useful for some purposes.

CHAPTER 20

TOPOGRAPHY OF OCEAN BOTTOMS

General Remarks

Tables 20.1 and 20.2 show the approximate areas of the ocean containing various depths of water. Table 20.3 shows the proportions of the Atlantic, Pacific and Indian Oceans which have depths exceeding 2,190 fm. The topography, or profile, of the ocean bottom is very irregular, there being submerged mountains which are termed RIDGES or RISES. Some of these rise precipitously for several thousand feet, and some, whose tops extend above water, form islands. The term SILL is applied to a submerged elevation separating two basins, such as the sea bottom of the Strait of Gibraltar. These irregularities in the ocean floor influence the movement of water masses, not only in the depths but near the surface, and thus effect the physical, chemical and biological properties of the water. Some elevations restrict the migrations of marine animals. Large depressions are called TROUGHS, TRENCHES or BASINS, and a DEEP is the lowest part of a depression. CANYONS are relatively long and narrow depressions in the continental shelf.

Table 20.1. The area of ocean occupied by various depths

Zone	Average depth (fathoms)	Area (millions of square miles)
Continental shelf (0–110 fm)	27	10·6
Continental slope (110–1,330 fm)	690	14·9
Deep-sea bottoms (1,330–3,140 fm)	2,415	109·5
Deeper depressions in deep-sea bottoms (3,140–5,740 fm)	3,335	4·3
Whole ocean area (0–5,740 fm)	2,080	139·3

Table 20.2. Areas and proportions of ocean which attain stated depths

Depths (fathoms)	Area (millions of square miles)	Percentage of total area
0–110	10·6	7·6
110–550	5·9	4·3
550–1,090	5·8	4·2
1,090–1,640	9·4	6·8
1,640–2,190	27·4	19·6
2,190–2,730	45·9	33·0
2,730–3,280	32·5	23·3
Over 3,280	1·7	1·2
All depths	139·2	100·0

To humans, the most important part of the ocean floor is the CONTINENTAL SHELF, which is conventionally defined as the underwater extension of the continents with depths of water up to 100 fm. Continental writers use the depth of 200 metres (109 fm) as its limit. It is, however, the change of slope and not the depth which decides where its edge lies, and any stated depth limit is therefore somewhat arbitrary. At its seaward edge the continental shelf drops rather abruptly into the depths, and the transition slope down to about 1,000 fm is usually well marked and called the CONTINENTAL SLOPE. In some places this change in slope occurs at about 55 fm; at others it may not be pronounced until depths of three times as much are reached. About 100 fm is usually the significant depth, and 50 miles is perhaps the average distance offshore at which it is reached. To the south of Newfoundland the continental shelf is 250 miles wide, and in the Arctic it is particularly broad, in some places wider than 300 miles. Off the British Isles this shelf varies in width from about 30 miles (off County Mayo in Ireland) to about 200 miles off Land's End.

On the American side of the Atlantic the continental shelf is usually much wider than on the east side, where it is narrow from Biscay to the Cape of Good Hope. In the Pacific the shelf is wide off Asia but narrow off the Americas, where there are depths down to 2,500 fm within 150 miles of the Chilean coastline.

On the continental shelf, relief often differs little from that of nearby land, but there are, however, some gullies of large proportions. There is a remarkable one (descending to 480 fm) in the Skagerrak and some off the coast of the U.S.A., with walls rising to 1,000 ft and some even to 6,000 ft above their valley floors. Over much of the ocean floor gentle gradients prevail, but great mountain formations are known to exist in the deeps. Some islands have slopes of as much as 35°; for instance, the upper base slopes of St. Helena and Tristan da Cunha.

Table 20.3. Percentages of the areas of the Atlantic, Pacific and Indian Oceans that cover the great depths of water stated

Depths greater than (fathoms)	Atlantic	Pacific	Indian
2,190	47·0	63·6	57·9
2,730	21·2	28·4	19·8
3,280	0·6	1·8	0·4
3,830	—	0·2	—

Atlantic Ocean

The average depth is about 2,130 fm; 2,020 in the North Atlantic and 2,190 in the South. The greatest known Atlantic depth is the Milwaukee Deep, 5,035 fm, in the Puerto-Rico Trench; another great depth is that of 4,565 fm, 60 miles north of the South Sandwich Islands. Depths exceeding 3,830 fm have been sounded in about lat. 27°N, long. 60°W, and on the equator in long. 18°W. There are 15 important deeps in the North Atlantic of 3,280 fm or more and six in the South Atlantic.

The Atlantic has two large depressions, one along its eastern and the other along its western side. These are separated by the S-shaped Mid-Atlantic Ridge, which runs from Iceland through the Azores, Saint Paul Rocks, Ascension, Tristan da Cunha, to Bouvet Island. On this ridge the depth rarely exceeds 2,190 fm, though in the depressions to the east and west of it depths of 3,280 fm are often reached, particularly in the western basin of the ocean. The ridge resembles a terrestrial mountain range in the convolutions of its crest (*see* Figs. 20.1 and 20.2). In places it is a cock's comb of parallel and very close ridges, with great depths and steep gradients in the narrows between.

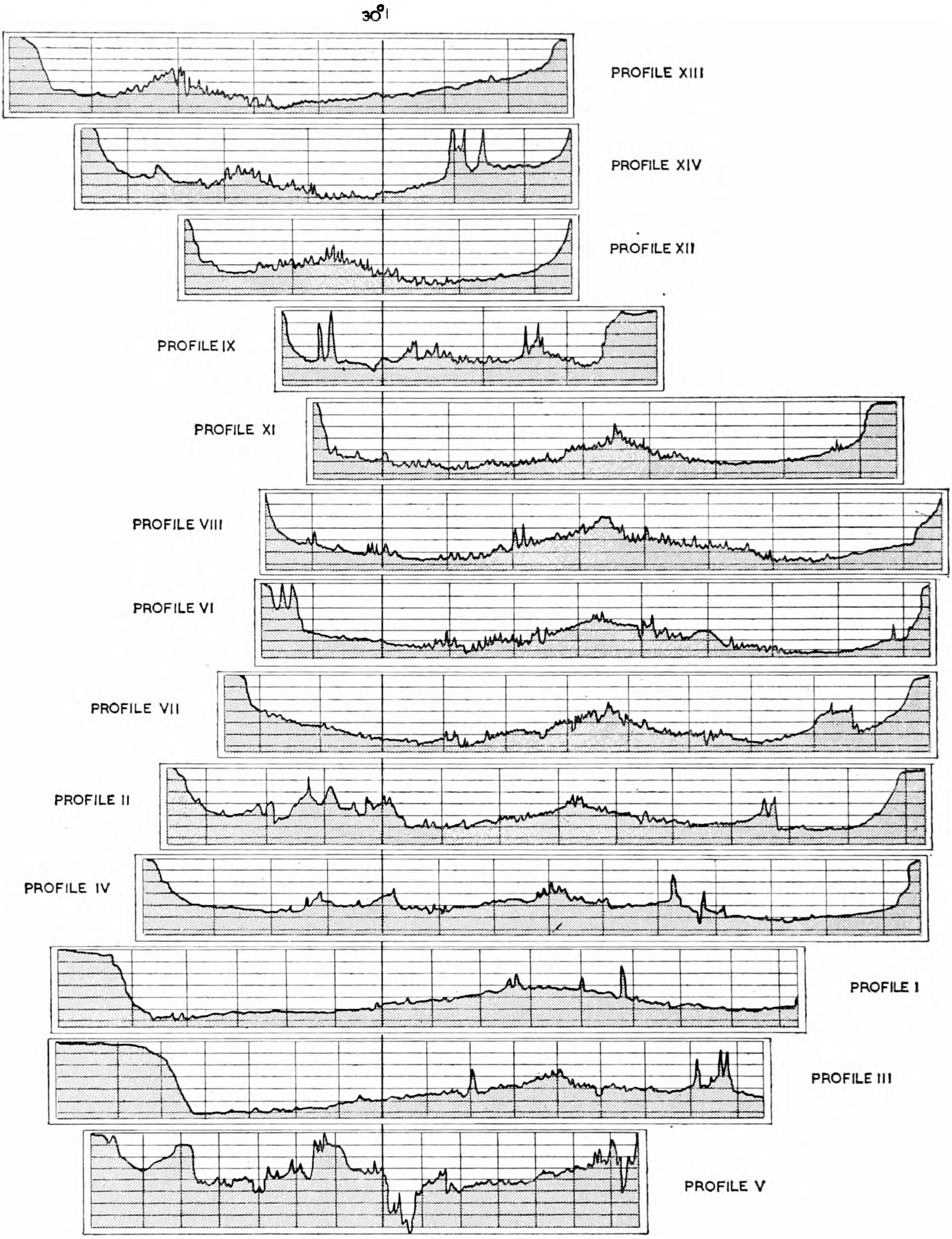


Fig. 20.1. The bottom topography of the Atlantic Ocean south of lat. 20°N
The location of each profile (section) is shown in Fig. 20.2. Note how clearly the Mid-Atlantic Ridge shows up in these echo soundings from the German *Meteor* Expedition, 1925–27. The scale for depths is marked for intervals of 1,000 metres (547 fm), and is exaggerated about 100 times with respect to the horizontal scale. The position of the meridian of 30°w is shown within each section.

At the equator in about long. 17°w, where the Mid-Atlantic Ridge trends east to west, there is a cut in it known as the Romanche Deep. This cleft of about 4,030 fm is of great oceanographical interest, since it connects the deep waters of the Brazilian and Sierra Leone troughs. The ridge broadens in various localities, as at the Telegraph Plateau (lat. 51°N) and the Azores Plateau; there are also lateral spurs from the ridge. Of these the most noteworthy is the Walvis Ridge, which trends NE to SW from Walvis Bay to Tristan da Cunha. Where this spur leaves the Mid-Atlantic Ridge, there is another one which trends west from the parent ridge and stretches towards South America to the Rio Grande. This spur is notched in lat. 40°S by a passage in which the water is 2,190 fm deep, and because of this the basin on the west side of the Atlantic offers freer passage to the northwards flow of the deepest waters than does the eastern basin obstructed by the Walvis Ridge.

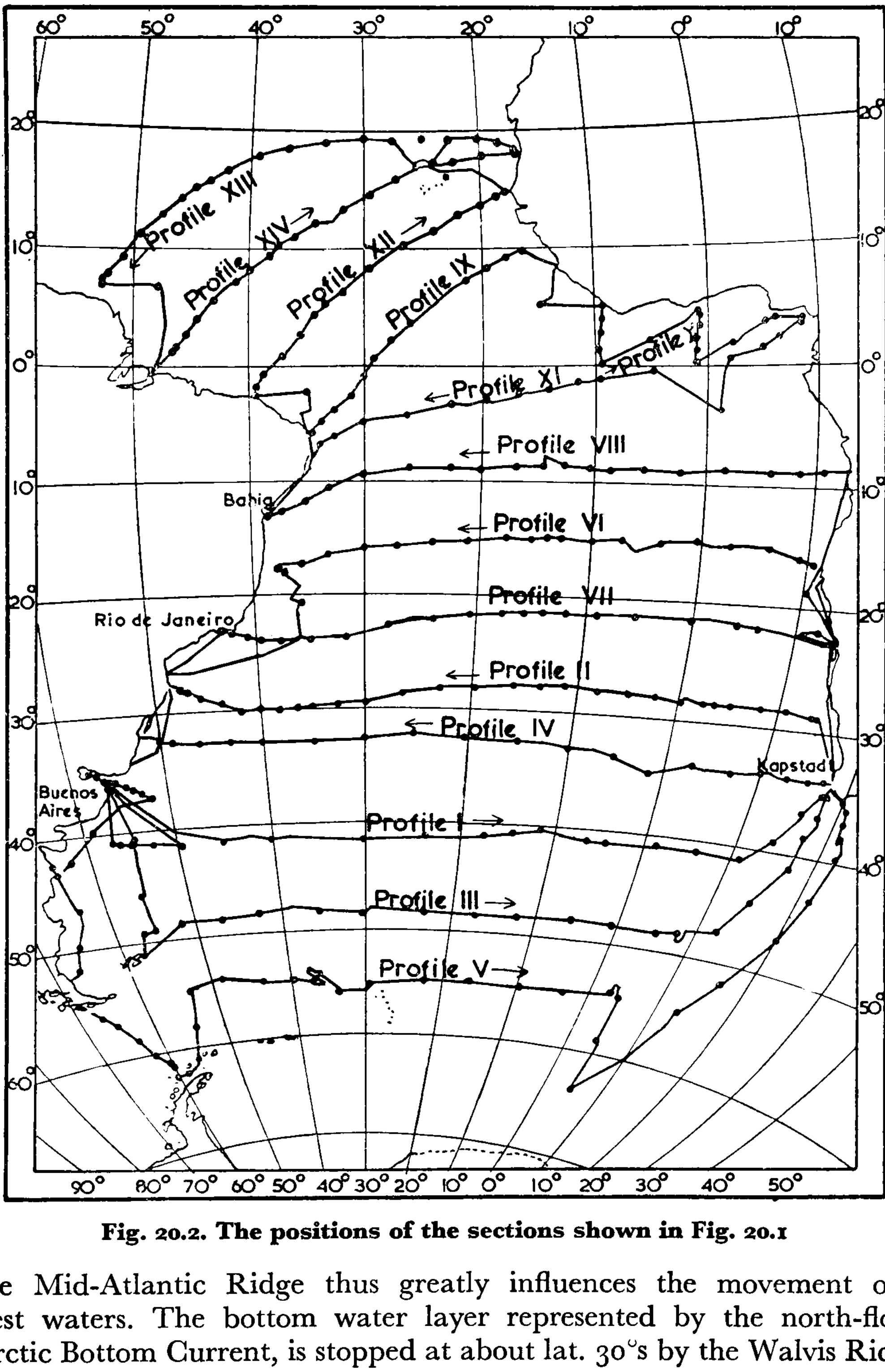


Fig. 20.2. The positions of the sections shown in Fig. 20.1

The Mid-Atlantic Ridge thus greatly influences the movement of the deepest waters. The bottom water layer represented by the north-flowing Antarctic Bottom Current, is stopped at about lat. 30°S by the Walvis Ridge in the eastern basin, but in the west it goes on further north. An overlying warm-water layer moving southwards as a compensating current, reaches correspondingly further south on the west side of the ocean than it does on the east.

Detail of bottom profiles shown by echo-sounding is interesting. A section of echo-sounding from the *Meteor* expedition is reproduced in Fig. 20.4 alongside one showing information derived from earlier lead soundings (Fig. 20.3).

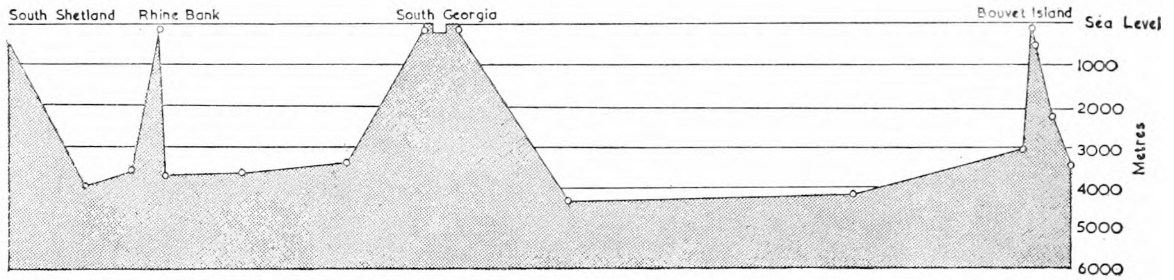


Fig. 20.3. The bottom profile along the line shown in Fig. 20.5 obtained by lead soundings previous to the 'Meteor Expedition'

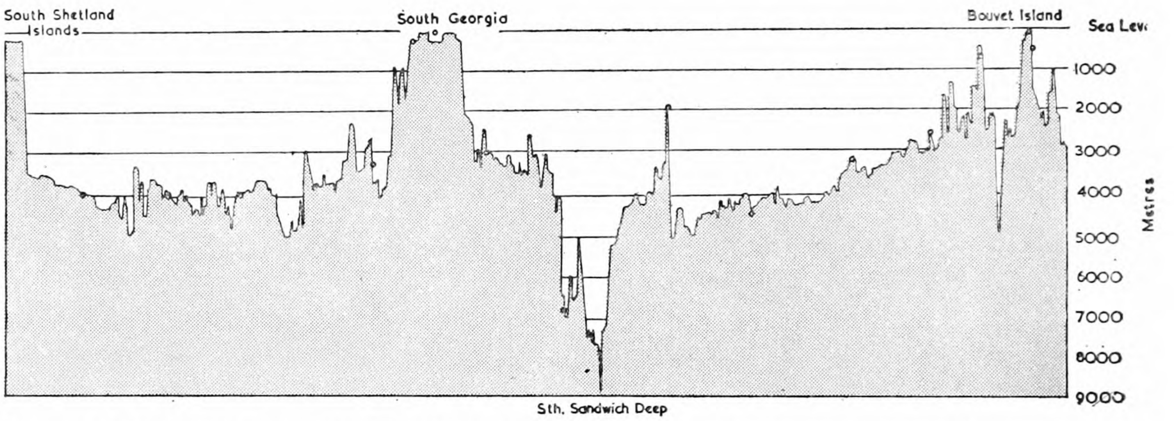


Fig. 20.4. The bottom profile along the line shown in Fig. 20.5 as obtained by the 'Meteor Expedition' from 2,485 echo soundings supplemented by 16 wire soundings

The vertical scale is exaggerated about 185 times.

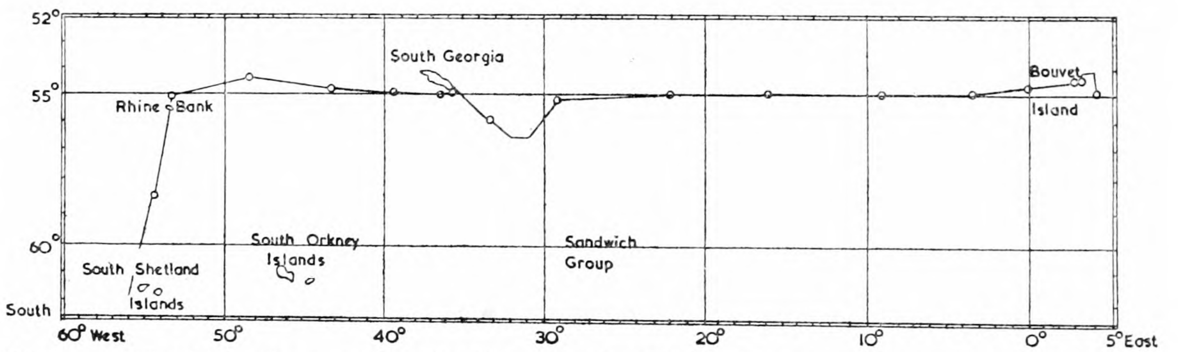


Fig. 20.5. The line along which the 'Meteor Expedition' made echo soundings between Bouvet Island and the South Shetland Islands

The small circles show the positions of the lead soundings which were previously available and from which Fig. 20.3 was constructed.

Indian Ocean

The average depth, as in the Atlantic, is about 2,130 fm. The greatest depths occur in the eastern part: 4,100 fm about 120 miles south of Java and 3,554 fm south of the Cocos Islands. The Gulf of Oman and the Bay of Bengal have less than 2,190 fm.

Along the Red Sea axis the depth is generally less than 550 fm, but there are deeps down to 1,310 fm. The Gulf of Suez is not deeper than about 44 fm. The Strait of Bāb-al-Mandab, between the Red Sea and Indian Ocean, has depths from 110 to 165 fm, but in the Gulf of Aden depths of more than 1,090 fm exist, with a maximum of 1,800 fm north-west of Socotra. In the Persian Gulf about 49 fm is the deepest sounding. The Indian Ocean has some 13 deeps sounding 2,730 fm and more.

In this ocean also there is a median ridge (the Mid-Indian Ridge) similar to that found in the Atlantic. Southward of lat. 20°S it trends south-easterly; north of that latitude it trends northward and north-westward towards Africa, and is known as the Carlsberg Ridge. North of lat. 10°N it is more or less on the meridian of 60°E and is there called the Murray Ridge.

Pacific Ocean

This, the largest of the oceans, accounting for about half the water area of the world, is also the deepest, with an average depth estimated at 2,240 fm (about $2\frac{1}{2}$ statute miles). It contains the maximum known depth of nearly 6,000 fm (*see* page 247) and there are many depths of more than 3,280 fm. The Pacific is said to have 33 deeps with 3,000 fm of water, against 19 in the Atlantic and 5 in the Indian Ocean. The Pacific contains water equal in volume to between five and six times that of all the world's land lying above sea level. Over about one-third of its extent the Pacific Ocean is deeper than 2,730 fm (*see* Table 20.3).

Arctic Ocean

This ocean is shallow to the north of Bering Strait, along the Siberian coasts and between Svalbard, Nóvaya Zemlyá and Franz Josef Land. Between the coasts of Norway, Iceland, Greenland and Svalbard there are two large basins with maximum depths up to 1,969 fm, separated by a rise on which depths of from 1,258 to 1,476 fm are found. There is a ridge, at places with less than 550 fm, between Norway, the Færöes, Iceland and Greenland.

In the polar region there is an immense depression, within which a sounding of 2,953 fm has been obtained (lat. 78°N, long. 175°W). At the North Pole itself, Papanin found the depth of 2,346 fm.

Antarctic Ocean

Over most of this ocean depths in excess of 2,190 fm occur, and near the South Sandwich Islands there is a trench with soundings of more than 4,375 fm. Depths exceeding 2,735 fm exist north of Enderby Land and to the north-west of Peter I Island.

CHAPTER 21

NATURE OF OCEAN BOTTOMS

Methods of Obtaining Samples of Bottom Deposits

The floor of the oceans is covered by deposits, the composition of which largely depends on the distance from dry land and the depth of water. Many kinds of instruments for bringing up samples of bottom deposits are in use and they have recently been much improved, so that it is now possible to take long sediment cores at great depths. A completely satisfactory tool for use on a vessel moving at 10 knots or more has, however, not yet been designed. Many types of dredges and grabs exist, amongst them sediment samplers which meet the needs of fishery-research workers interested in fish food on the sea floor. Some of the grabs are operated by pulling on light cables so as to penetrate the sea bed after the instrument reaches it. Others are designed to avoid displacement and distortion of the collected sediment.

In recent years many photographs have been taken of the sea bottom in depths up to 2,400 ft. Pictures of pronounced sand ripples at 500 ft depth are very striking, as also are photographs of starfish and other animals at depths of more than 200 ft. It is now also possible to photograph by simple instruments, standing upon the sea floor, the direction and rate of currents at great depths, which flow close to the sea bed. During the *Challenger* expedition (1873–76), sampling tubes were able to secure cores from 1 to 2 ft in length. The German *Meteor* expedition (1925–27) obtained cores up to 1 metre ($3\frac{1}{3}$ ft), and later an American scientist, using an explosive charge, was able to extend them to about 3 metres (9·8 ft). In 1942 Swedish oceanographers obtained cores up to 14 metres (46 ft) in length, and three years later one of nearly 70 ft was drawn up by their 'Piston Core Sampler'. With this instrument they have recently procured cores 20 to 50 ft long in the Mediterranean in depths from 1,000 to 2,000 fm.

By using explosive charges, modern oceanographers have gone far in determining the sediment thickness overlying the hard crust of the submarine lithosphere. In a depth of 2,000 fm in the Tyrrhenian Sea, a sediment thickness of 9,000 ft was found by the Swedish *Albatross* expedition; off the African coast the sediments were less thick. They discovered that only a thin layer of sediment overlies the hard crust in the Bay of Biscay and, in crossing the Atlantic, determined thicknesses of sediment 'carpet' of from 1 ft to 5,000 ft. The deep bottom of the Atlantic was found to be so uneven and to have so few plane surfaces that it is difficult in many places to interpret the results. In the Caribbean the bottom is much smoother.

Nature of Bottom Deposits

Close to shore, and on the continental shelf, marine deposits generally consist of sand, gravel and mud, with which the remains of animals and plants are mixed. These are usually found in depths of less than 100 fm and within an average distance of 75 miles from land.

Outside this zone, up to about 200 miles from land, the deposits are still influenced by the nature of adjacent coasts. In these deeper areas down to about 1,000 fm, i.e. still on the continental slope, they consist principally of blue, green and red muds, formed by the deposition of finely divided clays brought to the sea by rivers. In these deposits the shells of marine animals are found, and the different colours result from the presence of various minerals. Fine bluish-black or blue muds are commonest, though green mud occurs frequently; red muds are relatively rare.

In oceanic areas remote from land the deposits are chiefly the remains of organisms which live in the upper levels of the ocean, and in them are found small quantities of wind-carried volcanic debris and very fine terrigenous materials. The fine dust ejected by a volcano like Hekla in Iceland can travel very far by wind transport, certainly as far as Finland. The dust thrown out from Krakatau in Sunda Strait in 1883 circled the world. This volcano exploded on 26th and 27th August with what was alleged to be the loudest noise ever heard. All one day in the vicinity of the volcano it was as 'dark as the grave', owing to the ejected pumice stone and ashes. One ship reported meeting a bank of ashes which reduced her speed to ½ knot. Another ship got through the strait to Batavia covered with ashes, and reported having passed through sea areas covered with pumice 7 ft thick. Dust from the explosion remained in the upper air for several years, causing gorgeous red sunsets and other optical phenomena all over the world.

The sediment of the deep ocean bed also contains dust and matter from outside the world. The fine dust from the hundreds of millions of meteors which enter the earth's atmosphere daily is continually working its way to the earth, and much of it reaches the ocean bed. Since the oceans occupy nearly three-quarters of the earth's surface, the majority of meteorites which reach the earth in solid form also fall therein.

Our earliest knowledge of deep-sea deposits was mainly due to the *Challenger* expedition (1873–76), whereby a classification was built up distinguishing LITTORAL DEPOSITS between high-water and low-water marks, the SHALLOW-WATER DEPOSITS between low-water mark and 100 fm and DEEP-SEA DEPOSITS beyond 100 fm. There are two main groups: TERRIGENOUS DEPOSITS, formed in deep or shallow water close to land, and PELAGIC DEPOSITS (Greek *pelagos*, the sea), formed in deep waters far from land. A table was prepared showing the characteristic depth at which the more important deposits are found, the percentage content of lime (actually calcium carbonate) for each type and the area of ocean bottom covered by each. The lime content is important; Sir John Murray found that the proportion of calcium carbonate in sediments diminishes with increase in depth. There are, however, many exceptions to this rule.

If we imagine an isolated conical mountain in a tropical ocean, far from land, having its base below 4,375 fm of water and its summit rising to 275 to 330 fm, the general picture of ocean deposits would be as follows. From the summit of the mountain down to a depth of about 1,095 fm would be the domain of pteropod ooze, the sediments having a lime content of about 90%. Thence to about 1,640 fm would be the higher zone of the globigerina ooze, having a lime content of not more than 60%. Beneath this and down to about 2,735 fm would be the lower globigerina ooze zone containing some 20% of lime. Still deeper, in the regions carpeted by red clay (at depths around 3,830 fm), the lime content would have fallen to about 1%.

The smallness of the lime content in the deepest deposits is a feature which links up with bottom currents; indeed the distribution of the typical abyssal deposit, red clay, indicates the existence of the bottom current. The water of this current readily dissolves the calcium carbonate in the sediments over which it flows. The much greater latitude extent of red clay in the deep basin of the western South Atlantic than in the eastern, and the area of red clay in the deeps between the prime meridian and Angola, are explained by the existence of the clefts in the Rio Grande spur from the Mid-Atlantic Ridge and in the ridge itself, referred to on page 257. Further remarks on deep-sea deposits are given below.

PTEROPOD OOZE results from the sinking of delicate shells of tiny surface-living molluscs sometimes known as 'sea-butterflies'. This ooze is not found below 1,000 fm, but it covers 400,000 square miles of the Atlantic floor. It occurs on submarine ridges as far south as lat. 35°S, and is really a variety of globigerina ooze in which pteropods are plentiful. The commonest types of pteropods have sharply conical forms, but the shells ordinarily are broken. They seem easily soluble and are rarely observed in samples from deep water. The carbonate content ordinarily is high.

DIATOM OOZE is found at an average depth of 1,500 fm over a range from 600 to 2,000 fm, and occurs extensively in the Antarctic and North Pacific. It consists of the siliceous remains of a primitive group of microscopic plants, and with it an admixture of clay and radiolaria (*see* below) is commonly found. This ooze is white in colour, and a great belt of it round the Antarctic continent is found beneath the Antarctic Convergence (*see* page 225). This convergence is the meeting place of cold water of polar origin, usually heavily laden with diatoms, and the warmer water from temperate latitudes, laden with foraminifera. The result is a sinking of diatoms into the depths. The belt of diatom ooze encloses the glacial-marine sediments which exist northward of the Antarctic continent.

The sharp border-line between the hemipelagic sediments of the Newfoundland Banks and the deep-sea red clay of the North American Trough almost coincides with the corresponding Arctic Convergence in the northern Atlantic. There is thus an association between sediment zones and the meeting places of differing water masses.

GLOBIGERINA OOZE occurs at an average depth of 2,000 fm, mostly from 1,200 to 2,200 fm, though it has been found from 400 to 3,000 fm. It is the characteristic deposit in the Atlantic, and consists principally of minute calcareous shells of foraminiferal organisms, especially the characteristic globular form of globigerina. This ooze is dirty-white in colour when dried, and is of similar character to terrestrial chalk formations.

RADIOLARIAN OOZE is abundant in the Pacific, occurring at an average depth of 2,900 fm, mostly from 2,000 to 5,000 fm. It is not found in the Atlantic, and is nothing but red clay (*see* below) containing a large number of minute siliceous skeletons of radiolaria, which are a group of the protozoa. This ooze is often mixed with the main constituent of globigerina ooze, but the latter is rarely found in deposits in the greatest depths. Radiolarian ooze is red or dark brown and is less plastic than red clay alone.

RED CLAY. This deposit is the most characteristic and widely distributed oceanic deposit of all. It occurs in depths greater than 2,225 fm. It consists almost entirely of insoluble residues of wind-blown dust (generally of volcanic origin) which, by a long process of decomposition, form red clay. In the North Atlantic the colour is brick-red, owing to the presence of iron oxide. The

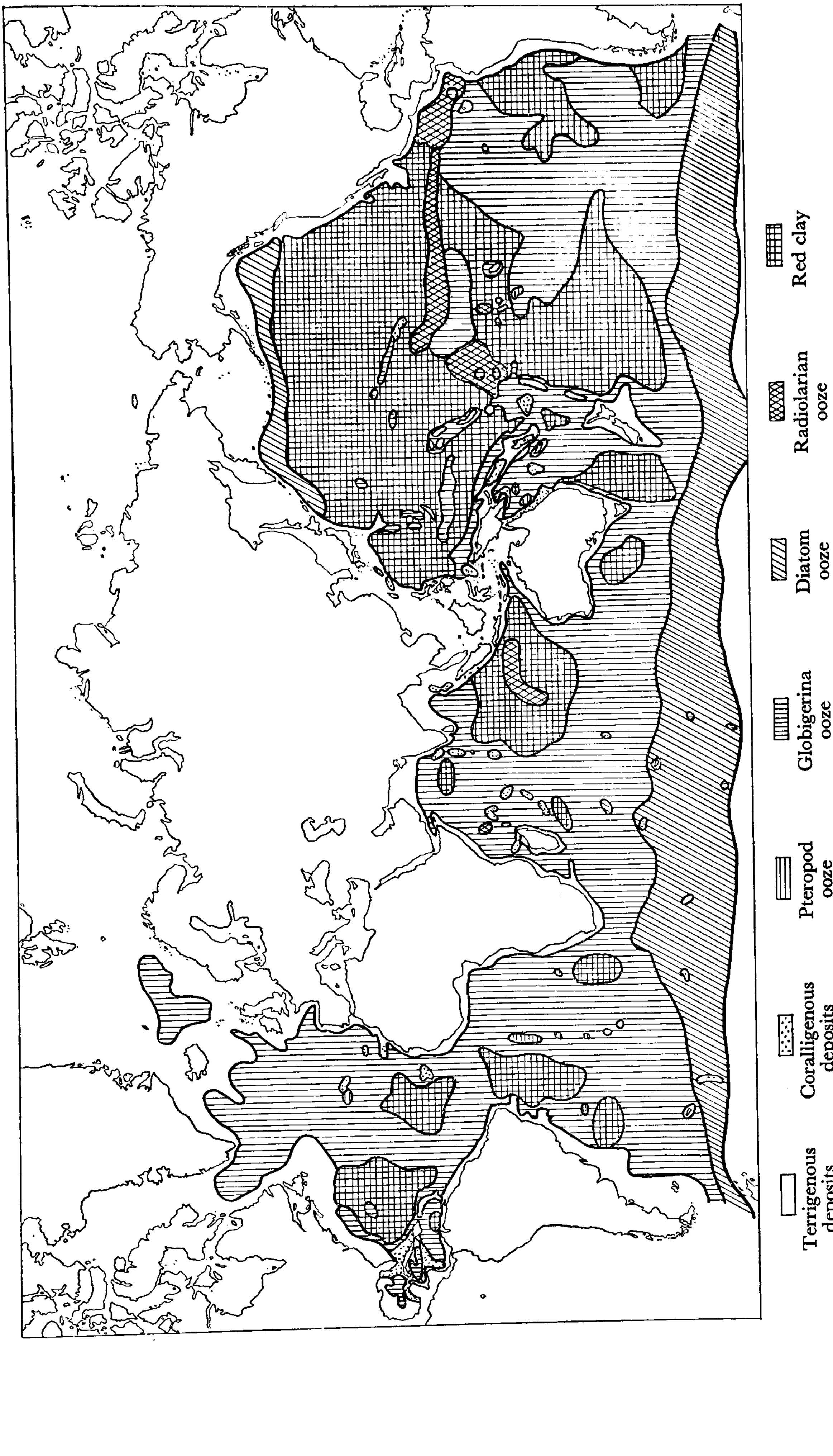


Fig. 21.1. Distribution of various types of sediment on the floor of the oceans (small areas omitted)

calcareous, and to a large extent the siliceous materials, which characterise the oozes, are generally dissolved as they sink slowly to the ocean bed; they are therefore rarely present in the greatest depths. Very fine clay particles might take 15,000 years to sink from the surface to the greatest depths, but they do reach the bottom ultimately.

Other deposits are known, such as CORAL MUD and CORAL SAND.

The German authority W. Schott has computed that blue mud accumulates in the Atlantic at the rate of about 1 in. in 1,400 years, globigerina ooze at the rate of about 1 in. in 2,100 years, whilst for red clay the figure is about 1 in. in 2,850 years. A Swedish expert, from studies of long cores, has, however, derived a settlement rate of 1 in. in 3,570 years for red clay in the Atlantic; for the central Pacific he estimates a rate seven times as slow. In the latter case, a sediment core 1 metre ($3\frac{1}{3}$ ft long) would indicate a time of about 1,000,000 years. Cores, 30 to 50 ft in length, taken from the Atlantic bed have been estimated to reach back in time for 1–2,000,000 years. A sediment 5,000 ft thick which is known to exist on the bottom of the Atlantic corresponds to some 200,000,000 years of uninterrupted deposition.

From the study of ash layers in core-samples, light is thrown upon volcanic activity in ages past. By investigating the content of the virtually indestructible pollen grains often found in sediments we can add to our knowledge of the geological history of the oceans.

A general idea of the distribution of ocean deposits is shown in Fig. 21.1.

Bottom Deposits and Oil

The most widely held view as to the origin of oil deposits ascribes the oil to plankton of ages long ago, plankton being the name given to all the minute freely drifting animal and plant life in the sea. The oil is believed to have been pressed out from the bodies of myriads of such tiny animals, once inhabiting the upper waters of an ocean where dry land now is. They died and settled to the ocean bed over great periods of time. The resulting ooze became overlain with other deposits as the geography of the ocean changed, and eventually there came into existence a highly compressed ooze sandwiched into sedimentary strata from above which the ocean had receded.

CHAPTER 22

INSOLATION AND LIGHT PENETRATION

Insolation

The oceans receive energy by short-wave radiation from the sun and sky, and lose energy by outgoing long-wave radiation, and on the average for all oceans the amount of radiation received is greater than that lost. Some of the heat not lost by radiation is used in the process of evaporation, which, for all oceans between lat. 70°N and 70°S, is estimated as the annual equivalent of a layer of water about 40 in. thick. This agrees with the probable value of average precipitation (a layer about 31 in. thick) combined with the estimated amount of water entering the oceans from the rivers (a layer about 10 in. thick).

Of the diffuse radiation from sky and clouds, Sverdrup states that only about 8% is reflected from the sea surface, so that on an overcast day 92% of this radiation penetrates the sea surface. On a clear day the amount of incoming radiation absorbed by the sea depends on the altitude of the sun, as shown below:

Table 22.1. Amount of sun's radiation absorbed by the sea

Sun's altitude (°)	Percentage reflected	Percentage absorbed
50–90	3	97
30	6	94
20	12	88
10	25	75
5	40	60

The net loss of heat by long-wave radiation from the sea depends on the temperature of the sea surface and the amount of long-wave radiation coming in.

Light Penetration

Ordinary daylight is composed of lights of different colours. These, when in the proportions found in daylight and reflected unchanged to our eyes, produce the sensation of whiteness. A white body reflects the same proportion of each of the different components and so appears white, but if illuminated by red light it looks red since it can then reflect red alone.

The colour of light depends upon its wavelength. White light separated into its constituent colours by a prism or other means, shows as a band of colour, red at one end and violet at the other, the red rays having the longest wavelengths and the violet rays the shortest. The wavelengths of light are measured in units, tenths of one-millionth part of a millimetre called Ångstrom units (Å) (10^{-8} cm). The wavelength so expressed is from 7,600–6,200Å for red light to 4,500Å for blue and 4,000Å for violet. As white light penetrates pure water, its constituent wavelengths are absorbed at different rates. The colours towards the red end of the spectrum are more rapidly reduced in intensity than those towards the blue.

In deep water, when no reflection from the bottom occurs, the light continues its downward journey, losing progressively its red, yellow and green components, until finally mainly blue rays are left. Some of the light on its downward passage, however, is scattered upwards to the eye, but these emergent rays will have had their blueness still further selectively intensified at the expense of other colours, because of the further removal of the latter during the upward passage. Thus it is mostly blue (with some green) light which survives to appear above the water surface, so giving the sea its typical colour.

The infra-red part of solar radiation, having longer wavelength than red light radiation, is absorbed in the top layer of water, only a fraction of an inch thick. This is made use of in aerial photography of coastal regions when it is desired to show the edge of the sea very clearly.

Sea water is never wholly pure, however far from land. It always contains suspended matter, the amount of which is usually greater in the colder parts of the oceans, especially in coastal regions. The suspended particles reduce the transparency of the water by stopping or scattering the light rays. This is a second factor in giving light that has passed through sea water a colour different from that of ordinary daylight. Experiments show that in fairly transparent oceanic water about 65% of the incident light is absorbed in the first $3\frac{1}{3}$ ft (1 metre). Only 17% reaches the depth of $2\frac{1}{2}$ fm (5 metres), 3·7% reaches to 11 fm (20 metres) and about 0·3% and 0·006% to 27 and 55 fm (50 and 100 metres) respectively. In coastal waters only 0·5% penetrates to a depth of $5\frac{1}{2}$ fm (10 metres).

Much information as to the transparency of sea water has been obtained by use of the SECCHI-DISC. This is a white-painted circular metal plate about 1 ft in diameter. It is raised and lowered slowly in the sea until the greatest depth from which it can be seen has been found. This may be as little as $3\frac{1}{3}$ ft (1 metre) in turbid river water, $2\frac{1}{2}$ fm (5 metres) in fairly clear lake water, increasing for sea water from about 7 to 11 fm in the English Channel to 27 or even 36 fm in the clear waters of the open ocean; actually the greatest depth of visibility is found in the Sargasso Sea. Apparatus using photo-electric cells has been recently used, usually in pairs; one instrument measures light intensity at the various levels in the sea, while the other on deck measures the intensity of daylight. Colour filters are used to study the quality of light reaching various depths.

There is a celebrated grotto in the island of Capri into which can penetrate only light which has passed through water and been deprived of its red, orange and part of its green and violet components. The grotto is consequently illuminated by a magnificent blue light. After passage through a glass tube 56 ft long, filled with North Sea water, daylight was found to be similarly deprived of components from the red end of the spectrum; but, because of the lesser clarity of the water, the colour of the emergent light was nearer green than blue. A diver in the Mediterranean at a depth of 16 fm (30 metres) recorded that the water viewed in a horizontal direction changed from greyish-green to greenish-blue according as offshore or onshore water currents prevailed. The veer towards blue became much more marked the deeper down he went. At between 14 and $16\frac{1}{2}$ fm (25 and 30 metres) red animals appeared black, and green or blue-green seaweeds appeared lighter in comparison.

The descents of Dr. William Beebe in his 'bathysphere', a steel sphere fitted with quartz windows, were of great value. He reached depths as great as half a mile. Beebe used a spectroscope to analyse the quality of the light at different depths, and his direct observations of the colour are in agreement with the general statements given above. He observed that animals had a colour complementary to that of the light in the depths. He noted a progressive loss of red, orange, yellow and green light with descent, and a complete absence of detectable light at 2,000 ft on his deepest dive on a bright day. The changing quality of the light with increasing depth was very impressive.

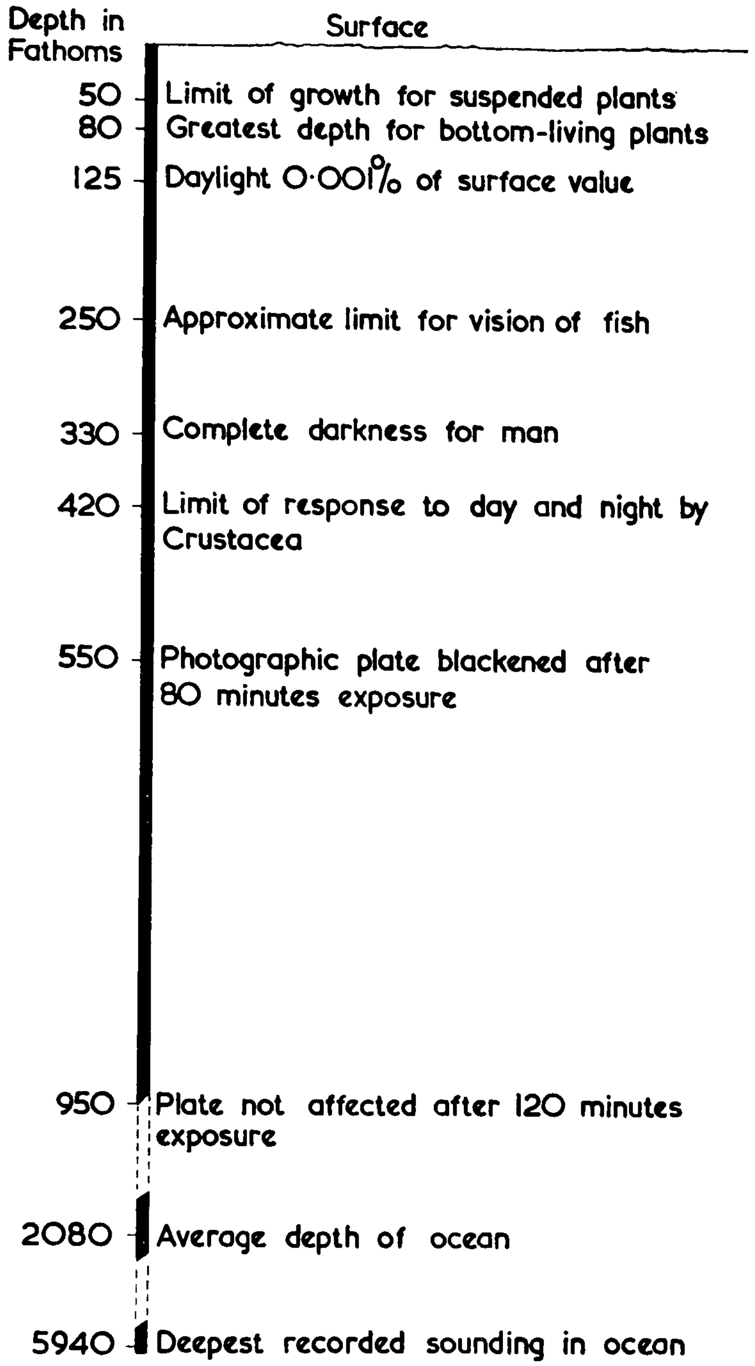


Fig. 22.1. The falling off with depth of the intensity of light penetration into the sea
The latest value for the deepest sounding is 5,980 fm, *see* page 247. The depth of ocean given is approximate.

The extent and quality of the penetration of light into the sea is of great importance to plant life in the ocean, and therefore the life of all sea creatures. The marine plants concerned are mostly microscopic in size, the freely floating diatoms, peridinians and flagellates. All plants require light to form their necessary sugars and starches from the carbon dioxide gas in the air or dissolved in the water. During this process free oxygen is liberated. It is well known that many plants which thrive in the open are unable to do so in deep shade. Moreover, for this process of PHOTOSYNTHESIS ('putting-together-by-means-of-light') the red and blue light is principally effective. Plants thus require an adequate amount of light of a suitable kind. Traces of light just sufficient to act upon a photographic plate are useless for this purpose (*see* Fig. 22.1).

The colour of seaweeds changes at different depths, in keeping with the above remarks. The brown seaweeds of the intertidal shallows merge with and are replaced by the red seaweeds as depth increases. Where the water is very clear these red weeds receive enough light for them to live down to depths of 71 fm (130 metres); at that depth their colour enables them to absorb the blue light which alone gets down so far. Such weeds in the deep water would appear black to a diver.

The amount of photosynthetic activity which occurs at different depths, as shown by oxygen production and carbon dioxide consumption, is one criterion of light penetration. At depths where there is much photosynthetic activity during daylight hours more oxygen is liberated than is consumed by the respiration of organisms, and with increasing depth a level is reached where these two quantities are just equal. This is called the 'compensation point' and is at the depth of 44 fm (80 metres) in the clear waters of the Sargasso Sea.

The amount of light in the different levels of the water controls what is known as the vertical migration of plankton. During daylight hours the tiny animals descend to a depth dependent on light intensity; at night they rise to the uppermost layers.

CHAPTER 23

MARINE BIOLOGY

Plankton

Plankton is the basic foodstuff of the sea. Without exception, animals must feed either on plants or on each other, living or dead. They are thus ultimately parasitic on plants, and would cease to exist were plant life to disappear. All plants need sunlight, so that we find the plants of the open sea in the upper layers of the water, where the sunlight is strong enough. They are not large, like the seaweeds of the shore, but are 'as a living dust scattered through the water'—to use the picturesque phrasing of Professor A. C. Hardy. The tiny floating plants, known as PHYTOPLANKTON, are fed upon by little animals (ZOOPLANKTON) scattered through the sea in teeming millions. Shrimplike creatures of many different kinds predominate, mostly varying in size from that of a pin's head to that of a grain of rice, but some are larger. Numerous other kinds of little creatures come under the term plankton, which includes all living organisms that drift about under the influence of the currents. The term is derived form the Greek *planktos* = wandering. Baby fish constitute part of the plankton and so do the early (larval) forms of many other organisms which, when adult, are large and can move at will, having then become able to compete with the currents.

Many fish such as herring, sprat and mackerel feed directly upon the little animals of the plankton. Most whales live directly upon slightly larger, but still quite small, prawn-like animals which swarm in the polar seas. These they are able to strain from the water by means of the great baleen plates in their mouths. It is strange that the blue whale, which is the biggest animal living (or which has ever lived) upon our earth, should find adequate sustenance from browsing directly upon what are more or less merely shrimps. It is, however, a different story with the toothed whales such as the sperm whale, which feeds upon large cuttlefish.

Only the freely drifting plants of the plankton (phytoplankton) live directly upon the mineral chemicals of the sea, and grow by using the energy of sunlight. Phytoplankton is most plentiful in spring, when the length of daylight increases and the water becomes more stable owing to the warming of the surface layers; downward convection currents are thus prevented, which might cause the vegetable plankton to be carried down to depths where the illumination would not be sufficient for it to continue living. The time comes when the phytoplankton, having used up the nutrient salts present in the upper illuminated layers, die away in large amounts. With the advent of autumn, the water layer starts to lose its stability due to cooling at the surface. Downward convection then takes place, and there is an eventual uprising of subsurface waters not deficient in nutrient salts; this accounts for another outburst of phytoplankton generally in autumn, though it is a smaller one than that of the spring.

Fish

An adequate account of the fish life of the sea would fill a volume. The herring, known as a 'pelagic' fish because it lives in the upper waters, is the object of an immensely important fishery. In British waters this culminates in the East Anglian autumn fishery, based upon Yarmouth and Lowestoft. The herring lays eggs (some 32,000) which do not float but attach themselves to irregularities of the sea-bed. The young forms which emerge from these eggs join the plankton, and the study of their fortunes is of importance to the fishery naturalist, as possibly deciding the proportion of the annual broods which survive to become adult fish.

That important food fish the cod lays 4–6,000,000 eggs which drift passively in the plankton, and fishes of this family can accomplish extensive journeys, as has been shown by marking them.

The flying fish, so remorselessly hunted by larger fish of the open ocean such as the dolphin, has been the subject of much recent research. Flying fish are said to 'fly' several hundred yards, but they do not really fly. They do not flap their wings but employ the principle of the glider. The flying fish swims rapidly to the surface, and, as it breaks through, sculls vigorously with its tail to become airborne with a speed said to reach 35 knots. Frequently, when it has nearly finished its flight, it will plunge into the crest of a wave and, by more vigorous tail-sculling, will take off again to glide a further distance. Wind-tunnel experiments with flying fish have been made using preserved specimens fitted with cellophane membranes in place of their shrunken natural ones. The results gave speeds in excess of 25 knots, starting with an initial speed of 40 knots. The flight durations were experimentally estimated at $1\frac{3}{4}$ seconds with ranges of 30 yd, but it was inferred that the better performances in nature would lead to glides averaging 2·6 seconds and to ranges of 40 to 50 yd.

The European freshwater eel has a remarkable story. This fish is widely distributed in the European rivers into which it originally makes its way in the 'elver' stage. Elvers are small eels which ascend into the rivers and water-courses in myriads. They assume eel form after having earlier been flat leaf-shaped creatures adrift in the Atlantic, and it has been proved that, three years before reaching our coasts, they originate from eggs spawned in the depths (500 fm and over) of a comparatively narrow area of ocean centred about midway between the Leeward Islands and Bermuda. Nobody has ever captured a perfectly ripe female eel, so that there can be no doubt that it spawns only in the sea. Whereas the European eel takes on eel shape and ascends our rivers after three years, the corresponding larval form of the American freshwater eel is ready for ascent into the rivers of the eastern United States after only one year. The larvae of the two forms originate in much the same area and are there mingled together in the water, yet the American form contrives in some mysterious way to separate itself from its European cousin.

Marine Mammals

As is well known, whales and their smaller relatives of the porpoise family are not fish but marine mammals. The whales' smaller cousins range in lengths from about 3 ft to over 15 ft and seem to enjoy following a ship or making spectacular leaps out of the bow waves. It is a matter of dispute how to allot the names 'porpoise' and 'dolphin', but many seafarers use the name 'porpoise' for all the small whale-like animals (which must come to the surface to breathe) and reserve the name 'dolphin' for that well-known blue-and-gold fish (a true fish) which is frequently to be seen darting about the ship near the surface in tropical regions.

Porpoises are not often seen singly, and may dive one after another in such perfect rhythm that, when seen from a distance, they have the appearance of one long undulating body. No doubt this is the basis of some sea-serpent stories. The flukes of the tails of these animals are parallel to the sea surface, whereas the tail fins of a true fish are up and down in the water. It has become evident of recent years that porpoises are, so to speak, 'fitted with asdics'; it seems beyond doubt that they can emit and receive supersonic signals.

Whales have to surface at intervals to exhale used air and inhale fresh. Their 'blowing' is the expulsion of impure air charged with water vapour, and the blow resembles a plume of steam. It can be seen up to 10 miles on a clear day, and the whales usually surface to blow five or six times at intervals of about 10 to 20 seconds. They then dive deep ('sound') for up to 20 minutes. Whales have been known to sound as deep as 400 fm when harpooned. When feeding deep, whale-bone whales may stay submerged for long periods; when travelling they usually stay near the surface. The maximum sounding depth of the sperm whale is an interesting subject for speculation; it feeds upon large cuttlefish, which live deep down in the ocean, and these must find their own food in the depths. The depth of the Subantarctic Intermediate Current, which leaves the surface in about lat. 45°s to flow northward up the Atlantic, attaining depths of from 270 to 550 fm (500 to 1,000 metres) in mid-latitudes, may afford a clue. Pelagic trawls at those levels have yielded quantities of deep-sea fish and crustacea. These may therefore well be the depths where the cuttlefish find their food (the crabs) and to which the sperm whales have to go down to find theirs (the cuttles). If this is so, the 'sounding depth' of sperm whales would be from one-third to two-thirds of a mile. New evidence (1957) from entanglement in submarine cables, and from the absence of the commonest pelagic squid in sperm whale stomachs, emphasises the great sounding depth of the animal.

The greatest of the whales, the blue whale, is a mammoth creature about which much is now known, thanks to the work of the Discovery Committee's expeditions. Their detailed investigations into the economics of the whaling industry, the food of whales and the chemistry and physics of the waters in which they live have been embodied in a series of reports. The blue whale averages about 90 ft in length when fully adult, and can weigh upwards of 100 tons, about equal to 25 elephants. The fin whale averages 75 ft in length. The flesh and the oils from this family of whales are edible, but those of the sperm whale (a member of a different family) are not generally considered edible.

Fin whales are generally seen in schools of 3 to 10, and have a maximum speed of about 15 knots. Blue whales are usually solitary (though sometimes in pairs) and travel easily at 10 knots, with the power to move at 20 knots for long periods if they so wish.

The 'whale catchers', which hunt the whales with explosive harpoons, are usually vessels of from 400 to 600 tons (140 to 160 ft long), steaming at from 12 to 16 knots. The factory ships have slipways up which whole whales can be drawn for cutting up and full treatment on board. The whales of the blue-whale family yield flesh claimed to be similar to beef.

During the 1938–39 Antarctic season 40,662 whales were killed and 2,887,832 barrels of oil were produced from them (a barrel contains one-sixth of a long ton). The taking of whales in the Antarctic is restricted by international convention, and the 1948 season was limited to four months with a permitted total catch of 16,000 'blue-whale units'. This means an overall limit of whales of various sorts equivalent to the taking of 16,000 blue whales. The products from the whaling industry are numerous, and the lubricating oil (consisting chiefly of waxes and not fats) obtained from the sperm whale is of special value, as also is ambergris, which comes from the intestines of the same whale. The depredations wrought by man upon the stocks of whales in the waters of the northern hemisphere and his inroads into the former stocks of walrus, seals and sea-cows is a sad but fascinating story. The walrus was not unknown in northernmost Britain in times past.

Other Marine Life

We have only remarked upon a few forms of marine life, but reference should not be omitted to one which connects up in a surprising way with wind-driven ocean currents. The Portuguese Man-of-War (*Physalia*), known to seamen as the 'Bluebottle', is a familiar and unpopular frequenter of Australian beaches, being carried there by warm currents from the tropics. It is often found drifting northward along the coast of Portugal, and is fairly common in the Mediterranean. It occurs in many other regions, but its sting seems not to have the same severity everywhere. Its most conspicuous feature is the bright-blue air bladder which floats on the surface of the water. This is broad and long, with an erect wrinkled crest along the upper surface, containing thirty or more 'glove-finger' chambers which keep it afloat unless the main air chamber is badly damaged. The whole may be inflated or deflated to maintain the right degree of buoyancy. Underneath the bladder is a mass of tentacles or feelers, some of which may reach depths up to 60 ft. If a fish strikes a tentacle it is almost instantly paralysed by the discharge of the stinging cells, and the sting can be dangerous to human beings.

Since the bladders or sails of these creatures are readily caught by the wind, they are widely distributed over the warm oceans of the world. A smaller type known as Velella, and to the seaman as 'By-the-Wind Sailor' or 'Sally Man', which has a more flattened kind of sail, may also be seen occasionally in tropical waters.

An American scientist has stated that the Portuguese Men-of-War of the southern hemisphere are of different physcial form from their fellows north of the Equator. In the former case they are so formed as to sail to the right of the wind, whereas they do the opposite in the northern hemisphere. This is a natural provision whereby these creatures, which have a deep 'grip' on the water, can counteract the wind-induced currents. Of hundreds of Velella recently seen in the Gulf of Mexico all were also sailing to the left of the wind.

Almost 200,000 species of animals are known to live in the sea, and 8,000 species of plants grow there in abundance. Evidence is accumulating that, acre for acre, the sea is more productive than the land. Suggestions have been made that man could sustain life by straining the plankton from the sea, and schemes for using plankton for human food have been put forward, but there are dangers from poisonous species, and recent American investigations resulted in the rejection of the idea. Nevertheless, in emergency plankton is of value as food and as a source of vitamins.

Deep-sea Life

From the world of planktonic life, dead and dying material is constantly sinking to the bottom. In the great depths of the oceans there is no light and therefore no true plants exist. The animals which live there prey upon one another or get their sustenance from the debris of dead animals and plants slowly raining down from the surface waters so far above them. They too are therefore dependent upon that world of light which is entirely unknown to them. Down at the very bottom are graceful 'sea lilies' and 'sea pens' with long stalks for anchoring themselves in the loose substratum. Down there, too, are relatives of the seaspiders of our own shores, but they are of relatively immense size and provided with long thin legs. It is a case of great size or of very small size with these bottom inhabitants of the deeps. The fish and, to a somewhat less degree, the squids and cuttlefish, which swim about in the continual search for prey in the deep and impenetrably dark waters, are the most curious and characteristic members of the abyssal fauna. The fish are usually coloured brown, black or silver, and are often of very odd shapes. Some of them have loosely fixed jaws and very distensible bodies, permitting the swallowing whole of prey bigger than themselves. From below 500 to 1,000 fm it is a general rule that the eyes of fish get larger; at greater depths they either are much larger still or become very small, or even disappear altogether. Many deep-sea animals have the power of emitting light (bioluminescence); some exude a luminous slime of red, yellow, green or blue, and one seaurchin is so brilliant as to have been described as a 'glory of the sea'.

There must be many animals living in the depths of the ocean of which we know nothing. Dr. Beebe speaks of many fish seen by him which had never been caught in nets, presumably on the score of their agility. Not many years ago, a specimen of a fish thought to have been long extinct (of the genus *Latimeria*) was caught off south-east Africa; this fish is known to have lived some 60,000,000 years ago, from its occurrence as a fossil in the chalk.*

Sea Coloration

The blue colour of the sea has been explained on page 267 as due to the selective absorption by sea water of light of other colours and also the scattering of blue light by the particles suspended in the water. The blue colour is most intense in the open oceans of middle and low latitudes, where it may be deep blue, brilliant ultramarine and sometimes, especially in middle southern latitudes, deep indigo.

In the open oceans of other latitudes and in shallower and coastal waters of all latitudes, the colour is modified by the presence of plankton. The plant constituents of the plankton produce a soluble yellow pigment. The combination of this yellow colour and the normal blue of the water produces the greenish tints of the sea which are observed, ranging from bluish-green to an intense green. As the quantity of plankton varies seasonally and at other times, according to conditions of food, temperature, etc., the extent and intensity of the green colour varies at different times and places. Because of the relative scarcity of plankton in the open waters of middle and low latitudes, where the sea is consequently bluest, it is said that "blue is the desert colour of the sea".

Another factor affects the colour of the sea in any latitude. The light by which we view the sea in daytime is composed not only of light retransmitted upwards from the upper layers of the water, but also reflected from the sea surface. This reflected light has naturally the prevailing colour of the sky at the time. If the

* Since this was written many more have been caught.

sky is blue, the blue colour of the sea is enhanced; if the sky is overcast or grey, the sea looks bluish-grey or at times even grey.

The colour of the sea also varies according to the direction viewed, relative to the sun's position at the time. The amount of sea disturbance also probably influences the colour. Sometimes streaks or patches of differing colours are seen, among which purple or violet tints may be noticed. Calcium carbonate in solution is believed to increase the blueness of the sea locally, e.g. in the neighbourhood of chalk cliffs or coral reefs.

Though, in a sense, the greenish colour of the sea due to vegetable pigment is a discoloration, the phenomenon known as 'discoloured sea' is a local one due to the actual colour in bulk of the animate matter suspended in it, or to any inanimate matter occurring locally. The colour of animate matter becomes visible when the plankton near the sea surface increases sufficiently. The most usual colours are shades of brownish-red, red and brownish-green, produced by one of two great groups of microscopic plants, the dinoflagellates and the diatoms. The dinoflagellates are simple little one-celled creatures possessing intricate cellulose cases, and two little hairs (called flagella) by the agitation of which they move. More rarely, other types of plant life increase sufficiently to produce other colour effects. Inshore discoloration is usually streaky owing to wave and tidal action.

Intense brown or red discoloration occurs at times in the Red Sea and in the Gulf of California (the 'Vermilion Sea' of the Spanish), hence the names given to these waters. This coloration, due to dinoflagellates, is at times very striking. The surface of the Red Sea has been known to look by day as though covered by floating reddish sand and, on collection, the colour was found to be due to myriads of tiny jellies about the size of a pin's head. These organisms, known as 'Noctiluca', had each a central spot of orange-red colour. Vast quantities of this organism are sometimes washed ashore so that the beach appears to be covered with piles of blood. Other localities where discoloration is especially frequent are the region of the Peru Current, South African waters and the Malabar coast of India. The red coloration in South African waters is called 'flower water'.

The seasonal increase of plankton normally occurs some time in the spring in frigid and temperate climates. On the Malabar coast it occurs in September or October, at the transition from the SW to the NE monsoon. Different organisms or stages of their development give a great variety of colour there, such as bluish-green, yellowish-green, amber, amber-brown, yellowish-red and red. Wide areas of the Arctic Ocean often become so deeply discoloured by dense growths of diatoms as to justify the name 'black water'. In the Antarctic, particularly in the South Georgia region, diatoms are so abundant in places as to discolour both water and ice floes over large areas. In the North Sea discoloration occurs which is known to fishermen as 'baccy-juice'.

Owing to changed water conditions, dense plankton growths may die off fairly suddenly, producing dirty-brown or grey-brown discoloration and 'stinking water', which may kill large quantities of fish. The Aguaje is a similar phenomenon on an extensive scale, occurring at times on the Peruvian coast, believed to result from an incursion of warm current from equatorial regions which kills all the plankton, fish or other living organisms of the cool Peru Current. Hydrogen sulphide is given off as the result of decomposition, hence the English name of the 'Painter' or 'Callao Painter', because of the darkening effect on ships' paint.

Larger masses of animate matter, such as fish spawn or floating kelp, may produce temporary discoloration.

Inanimate matter causing local sea discoloration may be mud from rivers, soil or sand particles carried out to sea by wind or duststorms or volcanic dust. In these cases the water is more or less muddy in appearance. The particles are mainly microscopic in size, though much larger than the fine dust which is partly responsible for the blue colour of the sea. Submarine earthquakes may also produce mud or sand discoloration in relatively shallow water, and oil has sometimes been seen to gush up.

Bioluminescence

This phenomenon, popularly known as PHOSPHORESCENCE, is very familiar to seamen. The light is produced by myriads of tiny animal organisms of the plankton and by marine bacteria. It is also produced by larger creatures, including many of the jellyfishes, brittle stars, crustaceans, bivalves and fishes. Though numerous marine animals are capable of this light emission, it is rare among shallow-water fishes, being mainly confined to those inhabiting the ocean depths.

When sea water is agitated, as in wave crests or the wake of a ship, the light produced may be bright enough to read by. The greenish glow, apparently from the water itself, is caused by innumerable microscopic organisms such as Ceratium or Noctiluca. Scattered throughout the luminous water are larger and brighter points of flashing light emitted by jellyfish and other larger creatures.

Bioluminescence is rare in land organisms, though the glow-worm and firefly are well-known examples. The light is now known to be produced during the oxidation of a chemical substance called 'luciferin'; as it is produced without emission of heat it is very efficient. Light without heat has never yet been achieved by human agency. The process has no connection with the very small amount of phosphorus present in sea water.

Many other forms of bioluminescence occur. Light of a more or less steady character may appear in an undisturbed sea. The commonest form is that of separate luminous bands or streaks, often of considerable length. In the form known as 'white water', the whole sea is of a uniform milky whiteness to the horizon and may give enough light to illuminate the clouds, if these are low. This phenomenon has been reported as having a calming effect upon the sea surface.

One of the rarest and most surprising forms is the PHOSPHORESCENT WHEEL, in which curved or straight bands of light diverge from a centre, like the spokes of a wheel. It is seen to rotate at such a speed that it cannot be an illusion due to the movement of the ship. The 'wheel' has been seen to fade and then brighten and begin to rotate in the opposite direction. Its cause is uncertain.

Bioluminescence at the sea surface is more common in tropical than in higher latitudes, although it may be very striking in the latter during the warmer seasons. It is particularly prevalent in the Arabian Sea, especially in July and August. The phosphorescent wheel has been observed in the Indian Ocean and China Sea.

Bioluminescence may reveal the position of ships to aircraft. During the war the Japanese manufactured a powder from a marine organism, to be moistened and rubbed in the hands so as to produce a faint light by which messages could be read.

Marine Growths

'Fouling' is due to the attachment and growth of plant and animal organisms on ships' bottoms, in water conduits and on piles or other submerged structures. Most materials suffer from these accretions, and the assorted organisms are popularly dubbed 'barnacles', 'coral', 'shells', 'seaweeds' or 'moss'. Barnacles or algae are usually the principal offenders, but a hundred or more different organisms may be present.

The fouling of ships is a matter of great importance in both the performance and the upkeep of ships. Ships may carry over 100 tons of fouling, after being a year afloat. After resting in the water for a few days the pontoons of flying-boats can acquire a load of several pounds of organisms which seriously affect their efficiency. The disadvantages of fouling in terms of speed, fuel costs, etc., are well known; hence the need for frequent dry-docking.

Ways and means of preventing fouling have taxed man's ingenuity since he first put to sea. The earliest mention of fouling is in connection with Remora, the fabled 'ship-stopper'. This comparatively small fish was mentioned by Aristotle in the fourth century B.C. Plutarch pointed out that growths were more likely to be the cause of loss of speed, and he stated that it was usual to clean the hulls of ships periodically. It is on record that in the time of Vasco da Gama (1469–1524) the Portuguese used to char the outer surface of the ship's hull to the depth of several inches. In 1720 the British built the hull of one ship, the *Royal William*, entirely with externally charred wood. In the reign of Henry VI it was stated that "in certaine partes of the ocean a kind of wormes is bredde which many times pearseth and eateth through the strongest oake that is".

Wooden sheathings over a layer of animal hair and tar were tried as well as lead sheathings, and the first record of copper sheathing relates to H.M.S. *Alarm* in 1761.

The prevention of fouling is a complicated problem and much effort has gone to its study. Oceanographical research has materially assisted, but more information is still needed about the organisms concerned. The problem is to prevent initial attachment of the organisms. A paint which continually exfoliates may be useful. Weight for weight, mercury is twice as effective as copper against barnacle fouling. There are many organic poisons which are more lethal than copper and mercury when in solution in sea water, but it is difficult to incorporate them effectively in paints.

Tropical harbours are usually more troublesome than those in temperate climates, and many marine fouling organisms can be shed in fresh water. The shells of barnacles and tubeworms can, however, remain attached to ships long after the animals are dead, but they become brittle.

Bacteria often assist fouling by providing a foothold for planktonic stages of fouling organisms. They may discolour bright surfaces, which are usually less prone to fouling than dark ones. Bacteria also attract other growths, being a source of food for barnacles, mussels, etc. They also assist the deposition of the limey cements of fouling organisms and may cause an increase in plant nutrients and so attract other unwelcome guests. Bacteria also form a protective layer over anti-fouling paints. On the other hand, some bacteria may release poisons under certain conditions.

The wood-borer Teredo, commonly known as the 'ship-worm', is a bivalve mollusc of the mussel and oyster family and can cause enormous damage. Shipworms 12 to 16 in long occur in our home waters; still larger forms exist in tropical waters. Passage into fresh water may kill the ship-worm, but unfortunately there is a form which lives in water more or less fresh. Each of the two valves of the creatures is divided into a pair of lobes with saw-like indentations. The boring is accomplished by rocking these lobes against each other, perhaps 8 to 10 times a minute. In 1503 Columbus was marooned for a year in Jamaica, due to the rotting of his caravels by this 'worm'. Protective measures are known, but efficient action against the creature calls for a good knowledge of its life history.

The common 'acorn barnacle' is the worse of the two barnacles, but occasionally the stalked or 'goose' form occurs. These latter may be attached to a hull by thick muscular stalks up to 3 to 4 in in length, at the ends of which are the bodies in their shells. Shellfish foulers are usually mussels and oysters, but these are easily poisoned and their presence on hulls denotes breakdown of the anti-fouling composition. Usually these shellfish foul underwater gratings most.

Tubeworms appear as white or greyish limey tubes which may be much coiled, lying flat against the surface or sometimes projecting outwards from it. They often occur in large patches on the hull, when they are commonly (but incorrectly) called 'coral' patches. They are very liable to foul propeller blades, and being capable of very firm attachment may occur almost along the whole length of the blades.

Sea as a Purifier

Indigenous bacteria are widely distributed throughout the sea and on the ocean floor; this results in such an effective decomposition of organic matter that the ocean has been described as the world's most efficient septic tank. Sea water, except near shore, is thus about as pure as any water in nature.

The pollution of sea water therefore presents a problem primarily on bathing beaches, in swimmimg pools and in connection with shellfish. It is often supposed that noxious materials passed into the sea become harmless quiet soon, and that dilution alone will render sewage innocuous. Experience with the sanitation of sea-water swimming pools shows that sea water is not entirely free from disease-causing organisms. Some of these are able to persist when others have been killed by chlorination. Completely effective purification has not yet been developed. Moreover, the chlorination of sea water presents unique problems owing to its salt content. Severe filtration of sea water can remove 'elements' which, left in, would destroy many noxious germs. Pathogenic bacteria can survive for a long time in modified sea water, and unless properly treated, pools filled with it may be more menacing than fresh-water ones. The water in fresh-water pools coming from municipal supplies is usually initially sanitary, but sea water may be pumped directly from a polluted bay.

Experiment has shown that 80% of the organisms in sewage were killed within half an hour in sea water and 99·9% after two days, but a few survived for nearly a month. Both enteric and coliform bacteria perish very soon in the sea. Their numbers decrease with distance from sewer outlets much more rapidly than can be attributed to dilution. In no case were the bacteria of cholera traced for more than a mile or two from the sewage outfalls into the open ocean; and only in solids or greases did they survive in the sea for long.

Control of malaria has been successfully undertaken in Albania and Jamaica by cutting channels to admit sea water into lagoons or swamps.

CHAPTER 24

RESEARCH VESSELS AND OCEANOGRAPHICAL OBSERVATIONS

Vessels

A typical modern research vessel, for long periods at sea, may be of about 2,000 tons displacement, up to 230 ft in length, have a crew of over 40 and carry six or more scientists. Such is the R.R.S. *Discovery*, famed for her work in the Southern Ocean. Research vessels engaged on fishery investigations are usually of the deep-draught trawler type, and are equipped with fishing nets as used commercially.

Biological Observations

The work is generally partly biological and partly devoted to a study of the chemical and physical qualities of the water and of the bottom. There will thus be a complete kit of nets, water-sampling apparatus and bottom-sampling gear. The nets are of many kinds. Large vertical nets of the finest silk are used for straining samples of the minute animal and vegetable life, including floating fish eggs, from the water. There are smaller silk nets for towing. These are streamed horizontally from a lead line, and some have devices so that they go down closed to a desired depth, fish at that depth only, and come up again closed. Some nets may have a small propeller to measure the amount of water filtered, and a depth-recording device whereby the same net can fish at various depths and give a record of each depth. There are dredges and grabs for taking samples from the sea-bed, and perhaps the coring instruments mentioned earlier.

Sampling Bottles and Reversing Thermometers

Various instruments are used for the non-biological work, including metal 'bottles' for taking sea-water samples from the depths. These bottles are sent down open to a desired depth as measured by a 'metre-wheel' over which a thin seven-stranded wire passes. When the desired depth is reached, a messenger sent down the wire closes the 'bottle' to imprison a sample of water within cylinders which insulate it from the other layers of water through which the instrument is pulled up to the surface. The bottle carries a very accurate thermometer with which to read the temperature of the sample. For work in very great depths, an insulating bottle cannot be used, for the sample on being raised to the surface would expand somewhat, with consequent fall of temperature. To avoid this, use is made of reversing 'bottles'; these are hollow metal samplers which carry 'reversing thermometers', resembling large-scale versions of the clinical thermometers used by doctors. When such a thermometer is turned upside down, the mercury column breaks at a constriction in the tube, and any temperature change on the way up to the surface cannot alter the reading shown. When a messenger is sent down the suspending wire, a reversing bottle, which was flooded and open, turns a complete somersault, so reversing an attached pair of thermometers and at the same time closing itself. This type of bottle is used down to the greatest depths, and six, seven or more of them are usually worked on one 'wire' at the same time. The closing of the uppermost bottle releases a second messenger just below itself, and this travels down to trip bottle number two, and so on down to the lowest bottle. It is customary with these bottles to use thermometers of two kinds: one which does not respond to pressure and one which does. 'Protected' and 'unprotected' are the terms used to name these. The pressure-resisting sheathing of a protected thermometer must not impair the ready conduction of heat to the thermometer bulb. The protecting outer tube is therefore partially filled with quicksilver and evacuated. The thermometer which contracts in volume* under the pressure of the head of water will read higher than the protected one, and from the observed difference in reading the depth can be ascertained. This technique, besides giving accurate depths of observations made on a slanted cable (due to ship drift), is used also to get soundings in great depths. Figs. 24.1 and 24.2 show bottles and Figs. 24.3 and 24.4 show thermometers of the kinds mentioned.

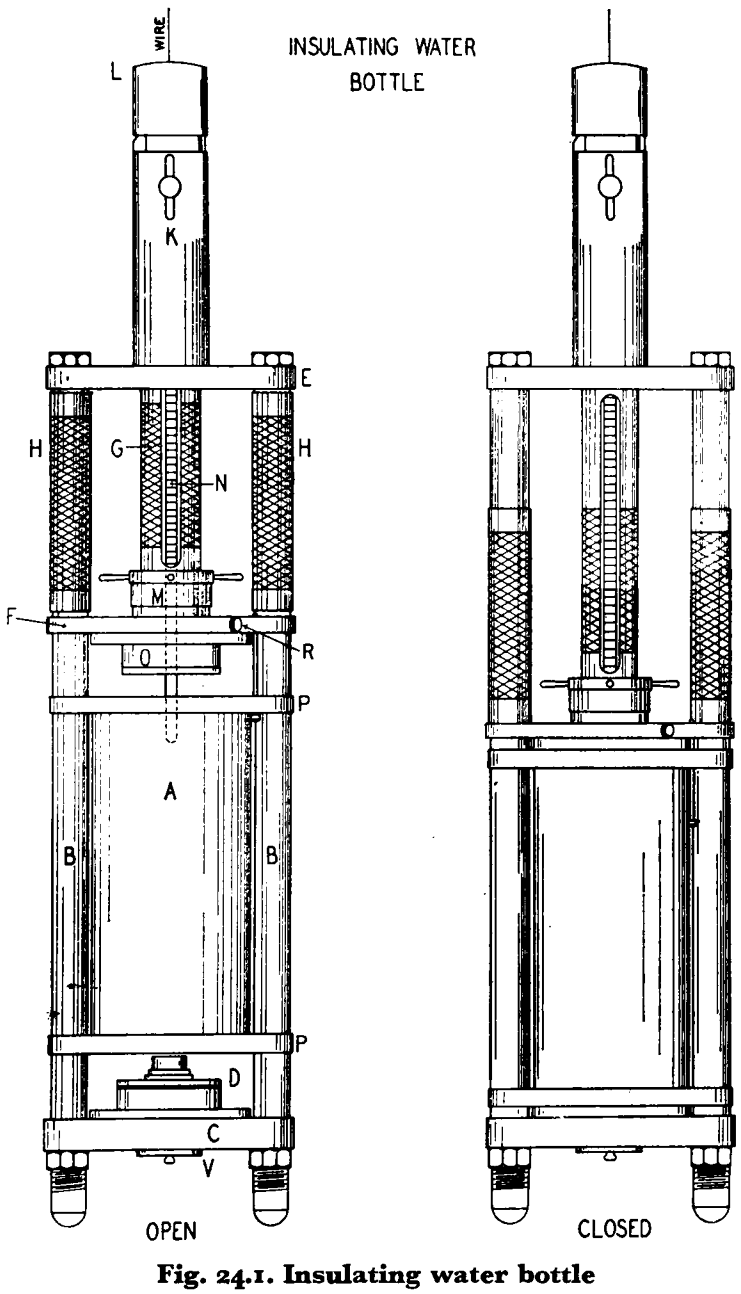


Fig. 24.1. Insulating water bottle

* The amount of contraction of the unprotected thermometers varies a little in practice but the effect averages 1 °C per 100 metres of depth.

M*

Variations of temperature at great depths are small and the reversing thermometers need to be of very great precision. To secure accurate data, a number of thermometers covering various ranges of temperature are used. It is important to locate any layer of water of less temperature than the surface water, but warmer than that immediately above and below it.

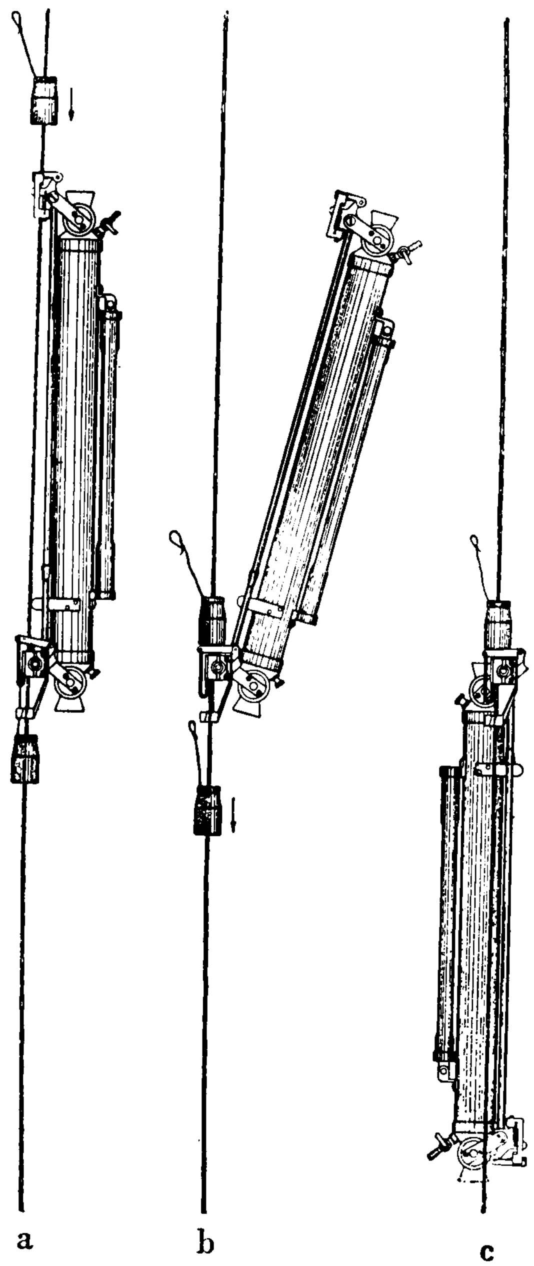


Fig. 24.2. Reversing water bottle

The bottle is shown open during the descent (*a*), in the act of reversing (*b*) and closed ready to be hauled up (*c*). Note the release of a second messenger which travels on further down the wire to trip a lower bottle.

Samples of sea water can be examined in the laboratory aboard the vessel; they are analysed for salinity, and the content of oxygen, phosphate, nitrate, silicate, etc. If the bacterial content is to be studied, or if the content of gold or other substances in the sea water is to be determined, bottles of special types are used. Large bottles may be used for the determination of the plankton content. The amount of vegetable plankton (phytoplankton) may be estimated from the density of coloration produced on releasing the green pigment from the plants by chemical means.

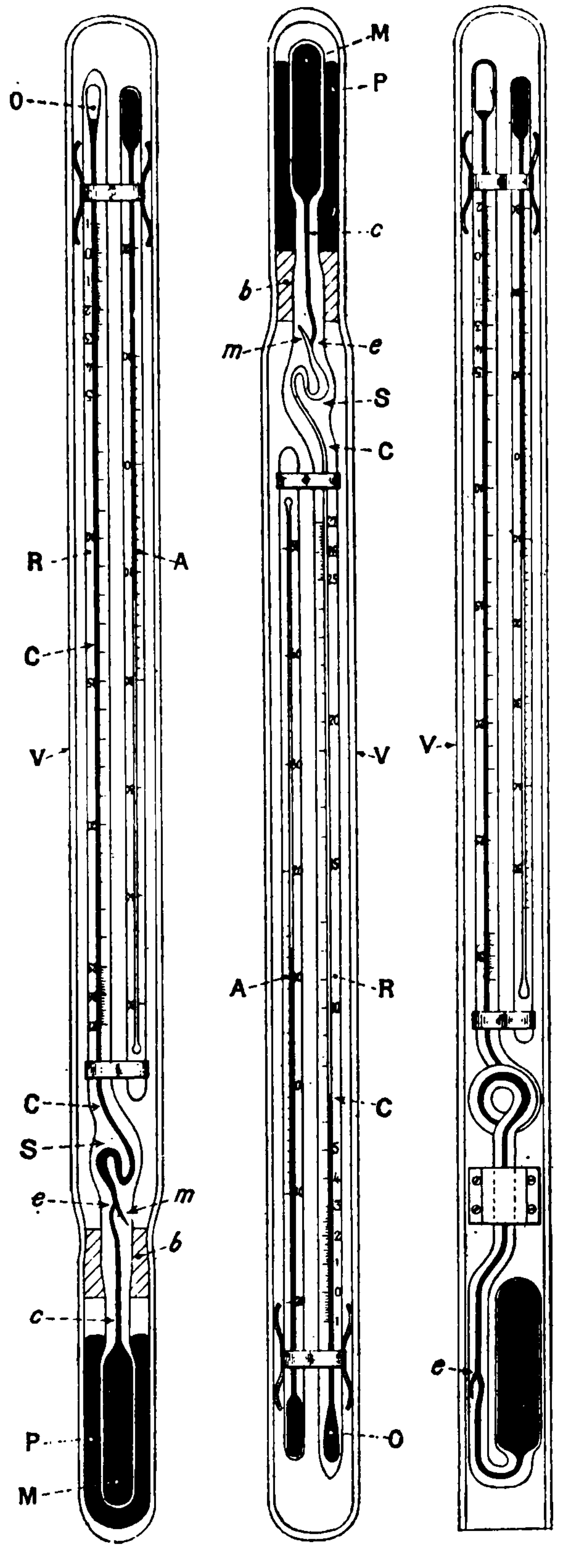


Fig. 24.3. Reversing thermometer

On the left is a reversing thermometer of the protected or pressure-resisting type shown upright as it is when sent down, and, in the middle, as it is after reversal ready for the temperature to be recorded. On the right is a reversing thermometer of the unprotected type, responsive to pressure; note its open end.

If, while on passage, water samples were needed for salinity a specially designed sampler might be used which, towed from the ship, contains the actual sampling bottle fitted with a thermometer which is read without withdrawal

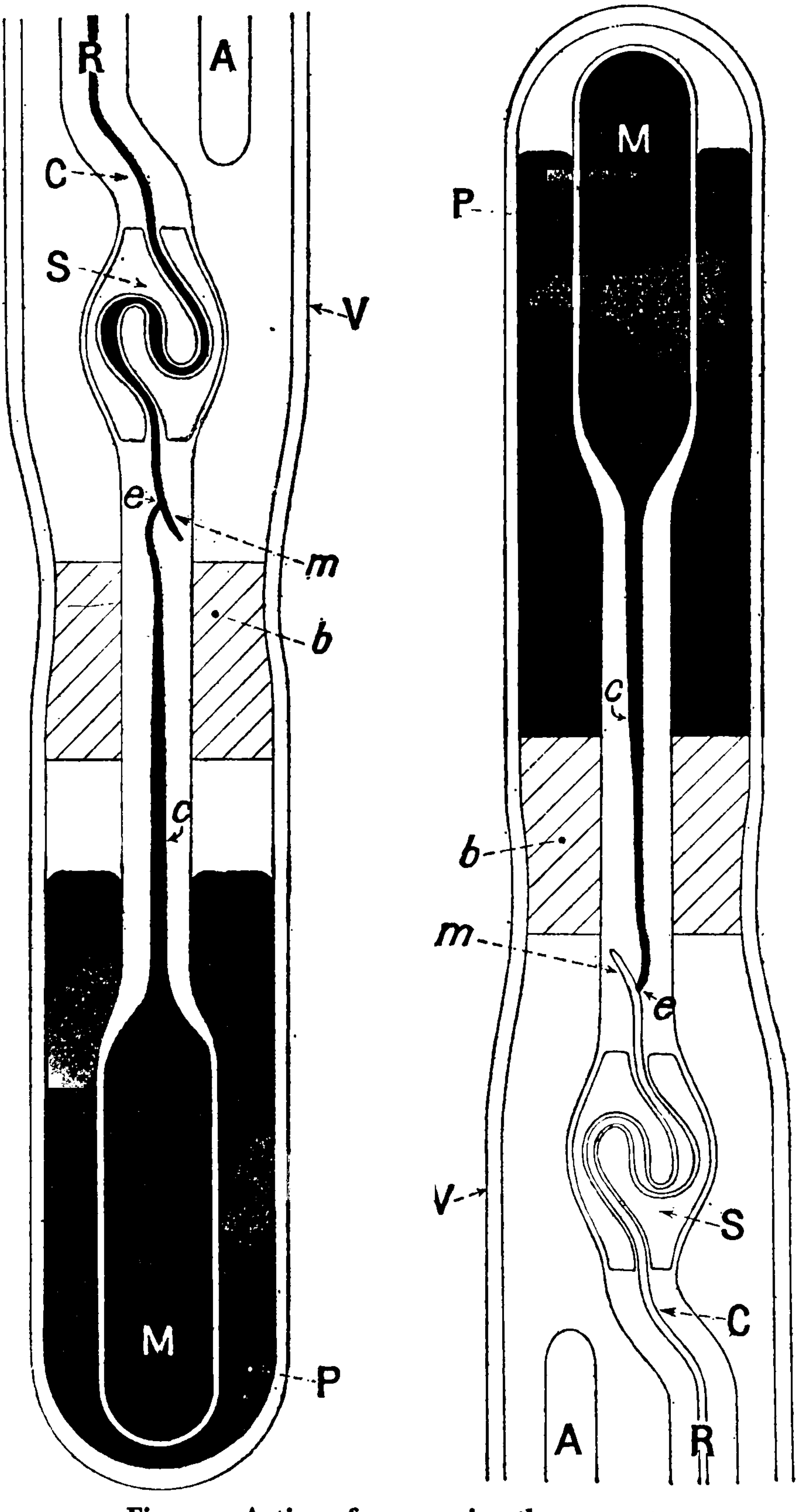


Fig. 24.4. Action of a reversing thermometer

Enlarged illustrations of part of a reversing thermometer to show the constriction at which the mercury thread breaks, before reversal (on the left) and after reversal (on the right). Note that if, through extra warming after reversal, the mercury expands downwards from the bulb, it is collected in the spiral and not allowed to fall down the graduated stem to falsify the reading.

from the sample. Salinity can also be determined by electrical means or by a refractometer, which measures the refraction of a ray of light passed through the sample.

There may be sieves and settling tubes with which to examine bottom samples, and the amount of radioactive matter in the sediments may also be studied. For problems of coast erosion and accretion, an instrument known as a sand-trap may be used to determine the quantity of silt or sand carried by the water at various depths.

Other Physical and Chemical Observations

The intensity of daylight above the water and at various depths can be simultaneously measured by photometers. Colour filters may be employed to ascertain which wavelengths of light penetrate to different depths. The transparency of the water can be investigated *in situ* at considerable depth by means of Secchi discs (*see* page 268). The rate of evaporation of the surface water can be found by noting the progressive increase of salinity in pans of sea water left exposed to the air on deck. The viscosity (internal friction) of water may be studied in connection with the vertical distribution of plankton. The viscosity of sea water is nearly twice as great at 41 °F as it is at 77 °F, so that tiny planktonic animals can remain poised in the colder water with less muscular effort than in the warmer water. Those inhabiting the warmest oceanic regions often possess bodies with feathery attachments so that they can remain more easily suspended.

Particular attention is given to those levels at which the salinity, temperature, density and oxygen content change most rapidly with moderate change of depth. These levels, known as 'discontinuity layers', are subject to marked changes of depth at different times. A useful instrument, known as the 'bathythermograph', provides a continuous trace connecting temperature and depth scratched by a sharp 'pen' on a smoked-glass slide, and permits close study of the microstructure of the water. It gives more information than sampling bottles can, and instruments are being brought into use to deal similarly with salinity. Another instrument, towed astern and connected to the ship with an electric cable, records temperature and salinity at the depth at which it is being towed.

Routine observations of sea surface temperatures would be made throughout the voyage, the samples being drawn up in a canvas or leather bucket, and the vessel might have a thermograph in the condenser intake, to give a continuous record of sea temperature changes.

Observations of electrical potential might also be made and the study of waves and of meteorology would also receive attention.

The hydrogen-ion concentration and the oxygen content of the water is studied, as these are of great importance in connection with the life processes of the minute animals and plants living in the ocean. The connected processes of assimilation and respiration leave their imprint upon the water masses. Sudden increases of growth of phytoplankton, which generally occur in spring, are related to high oxygen content in the upper water layers. Information about the speed of deep currents may also be obtained from the oxygen content. The waters of Antarctic origin flowing northward in the depths of the South Atlantic, carry an oxygen content which reflects the season at which they left the surface in the south. Lines of stations run in the north-south direction reveal oxygen maxima in the depths which imply a departure of the water from the surface in spring or summer. These maxima are therefore distant apart by a number of miles corresponding to one year's movement.

Currents

In moderate depths the research vessel might anchor to observe currents, and for this a variety of instruments is available. Some have propellers rotated by the current with a mechanism to record the number of rotations made, and a device whereby small balls are liberated from a hopper (proportional in number to propeller revolutions) and roll down a groove on the top of an arched compass needle into a box with radial compartments so as to indicate speed and direction. Other current-meters stamp direction and speed of water movement on to a thin metal strip, and simpler ones consist essentially of an open cylinder let down into the sea on a thin cable and designed to tilt a circular spirit-level clamped into a gimbal frame on the ship's rail.

The observations are usually repeated at many places over wide areas of an ocean. If a close network of 'stations' has been worked, it is possible to determine the flow of current at the surface, and at various depths, from the distribution of water densities.

CHAPTER 25

VARIATION OF TEMPERATURE AND SALINITY WITH DEPTH

Temperature

Broadly speaking, temperature decreases with depth in all oceans, but this is not always true in detail and there are no simple generalisations about variation of temperature with depth which will hold for all regions, or at all times.

We know, from the extensive use of deep-sea thermometers, that with descent into the depths at certain places there may be a fall of temperature for a time, then a rise, and then a progressive fall again. Warm water may, and often does, underlie cold water in areas where surface currents from the polar regions meet and spread over bodies of warmer and denser water (denser because much more salt) moving polewards from lower latitudes. Fig. 19.1 shows that warm sea water can be denser than cold if it is salt enough. This figure also shows that if two bodies of water equal in density, but of very different temperatures and salinities, come together and mix, the resulting mixture will be heavier (bulk for bulk) than the water of either of the original masses. For instance, water of salinity 30°/₀₀ and temperature 36°F (density 1,024) on mixing in equal amounts with water of salinity $36\frac{1}{2}$°/₀₀ and temperature 80°F (also of density 1,024) would produce water of salinity $33\frac{1}{4}$°/₀₀ and temperature 58°F, with a density greater than 1,024. It follows from this rather extravagant example that the mixed water would sink and might so spread out in the depths as to result in a temperature increase with depth being observed at one place and a decrease at another nearby location.

Hydrosphere

Just as the atmosphere is divided into troposphere and stratosphere, so oceanographers divide the ocean into two main divisions in depth, with the same names. The MARINE TROPOSPHERE is the upper layer of relatively high temperature which exists in middle and lower latitudes. In this layer, relatively strong currents are present and temperature as a whole falls off fairly rapidly with depth. In the MARINE STRATOSPHERE the salinity is lower and the temperature decreases very much more slowly with increasing depth; this part of the ocean includes the cold deep and bottom water, with a general arrangement into isothermal layers, extending to the ocean bed. The troposphere extends down to depths varying from about 270 fm to 550 fm. Between this and the stratosphere is a shallow discontinuity layer.

As stated in Chapter 19 the range of sea surface temperature over the open oceans is from about 29°F to 88°F. At 220 fm the range is from about 35·6°F to 62·6°F, and at 550 fm from about 33·8°F to 46·4°F. The bottom temperatures of the oceans range from about 32°F to 37°F or 39°F. Low temperatures are widespread at the ocean bottoms due to currents of cold water originating in polar regions and moving towards the equator. Certain of these deep currents have been estimated to move at very slow speeds, about 1 to 2 miles a day.

The following table, by Professor Defant, shows the broad structure of the temperatures in the troposphere and stratosphere of the Atlantic between lat. 50°N and 50°S. As explained above, this generalisation does not apply to every part of the region; for example, between lat. 20° and 30°S in the Atlantic Ocean a rise in temperature with depth is observed below 820 fm. This is due to the presence there of a tongue of colder water of southern origin overlying warmer and salter water which flows southward in the depths.

The quick decrease of temperature between 330 and 440 fm depth is very noteworthy. The large reduction in the rate of fall of temperature below 550 fm indicates that this is the boundary between troposphere and stratosphere.

The upper part of the marine troposphere is the region wherein daily and seasonal variations of temperature occur. Daily variations are found down to depths of 10 to 15 fm, and seasonal variations to about 10 times this depth. It is also the region where mixing of water occurs, owing to wind stress and to daily convection, in which process cooler water sinks, to be replaced by warmer water from below. All this tends to equalise the temperature of the upper part of the troposphere. As a result the temperature may be more or less uniform in this layer, or even show inversions. These remarks apply to the middle and lower latitudes of all oceans.

Table 25.1. Temperature variation with depth

Depth		Temperature (°F)	Rate of Temperature fall (°F) per 300 ft
(fathoms)	(metres)		
0	0	60·8	
			0·4
55	100	60·4	
			0·5
110	200	59·9	
			1·6
220	400	56·7	
			3·4
330	600	49·8	
			4·3
440	800	41·2	
			1·2
550	1,000	38·8	
			0·6
1,090	2,000	37·6	
			0·005
1,640	3,000	37·0	
			0·004
2,190	4,000	36·7	
			0·002
2,730	5,000	36·5	

Potential Temperature

Discrimination is made between water masses in the depths, having all characteristics save temperature more or less the same, by calculating what are known as potential temperatures, i.e. the temperatures which the water samples would possess if raised to the surface. This is done because deep water sinking still deeper into pronounced depressions on the ocean bed warms up somewhat on account of the extra compression. This is known as the ADIABATIC INCREASE. In the depths, water thus has a temperature *in situ* notably higher than it would have if a sample of it were raised to the surface. Samples of the same bottom water might therefore have different temperatures at different places, and errors in deducing the origin of the water masses might be made in the absence of temperature adjustments. The same body of water might, for example, change in temperature by about 0·45°F between 2,190 fm and 3,280 fm. The water *in situ* at a depth of about 2,190 fm in the eastern basin of the South Atlantic was stated by Wüst to have a temperature some $4\frac{1}{2}$°F higher than it would have if raised to the surface. These temperature increases were previously ascribed to earth heat or volcanic warming.

Salinity

The variation of salinity in the depths of open oceans is small, from about 34·6‰ to 35‰. This is because the waters there are so far removed from influences which can change their salinity. In the Atlantic between lat. 40°N and 40°S, salinity is highest at the surface, with a falling off (at first pronounced and then weak) to values of from 34·3 to 34·5‰ at 450 to 550 fm. There is a new increase to values from 34·8 to 34·9‰ at 875 to 1,100 fm. Thereafter increasing depth shows little change. This applies generally for latitudes between 40°N and 40°S, with the exception of a narrow equatorial belt in the Indian and Pacific Oceans, where salinity falls steadily from surface to bottom. Close to the equator the salinity maximum may be situated at around 50 to 100 fm, due to the copious rainfall reducing the salinity of the surface layers.

The vertical distribution of salinity in the oceans is governed by the circulation of the surface and subsurface waters. The widespread minima at 450 to 550 fm are ascribable to currents at intermediate depths, flowing towards the equator, which left the surface in sub-polar regions. The increase, in the layer below this, is due to currents flowing towards the poles from lower latitudes, bringing waters which left the surface in salter regions. A special feature of the Atlantic is the spreading-out at intermediate depths of the salt waters which emerge from the Mediterranean through the Strait of Gibraltar, below the surface current entering that sea.

The variation of salinity with depth is sometimes of economic value and the temperature difference between surface and near-bottom in the tropics has been used by French engineers to supply the motive power for experimental turbines, and is the basis of a scheme for ameliorating the climatic conditions of certain tropical cities.

Water Circulation in the Depths

The circulation in the deeps depends materially upon the sinking in polar regions and subsequent travel of cold heavy waters. In the South Atlantic there is free and wide communication with the Antarctic at depths greater than 2,190 fm, whereas in the north communication with the Arctic, besides being much narrower, is cut off by the bottom topography below depths of about 330 fm. In the Indian Ocean cold polar waters in the depths flow equatorward only from the Antarctic, apart from relatively small amounts originating in the Atlantic. In the Pacific the main communication is with the Antarctic; in comparison, that with the Arctic is negligible.

Figs. 25.1 and 25.2 show the vertical distribution of temperature and salinity in the oceans, and give some idea of the main features of the circulations in their depths. In the Atlantic the influence of a north-going flow of cold water on the bottom from the Antarctic is detectable as far north as the Bay of Biscay.

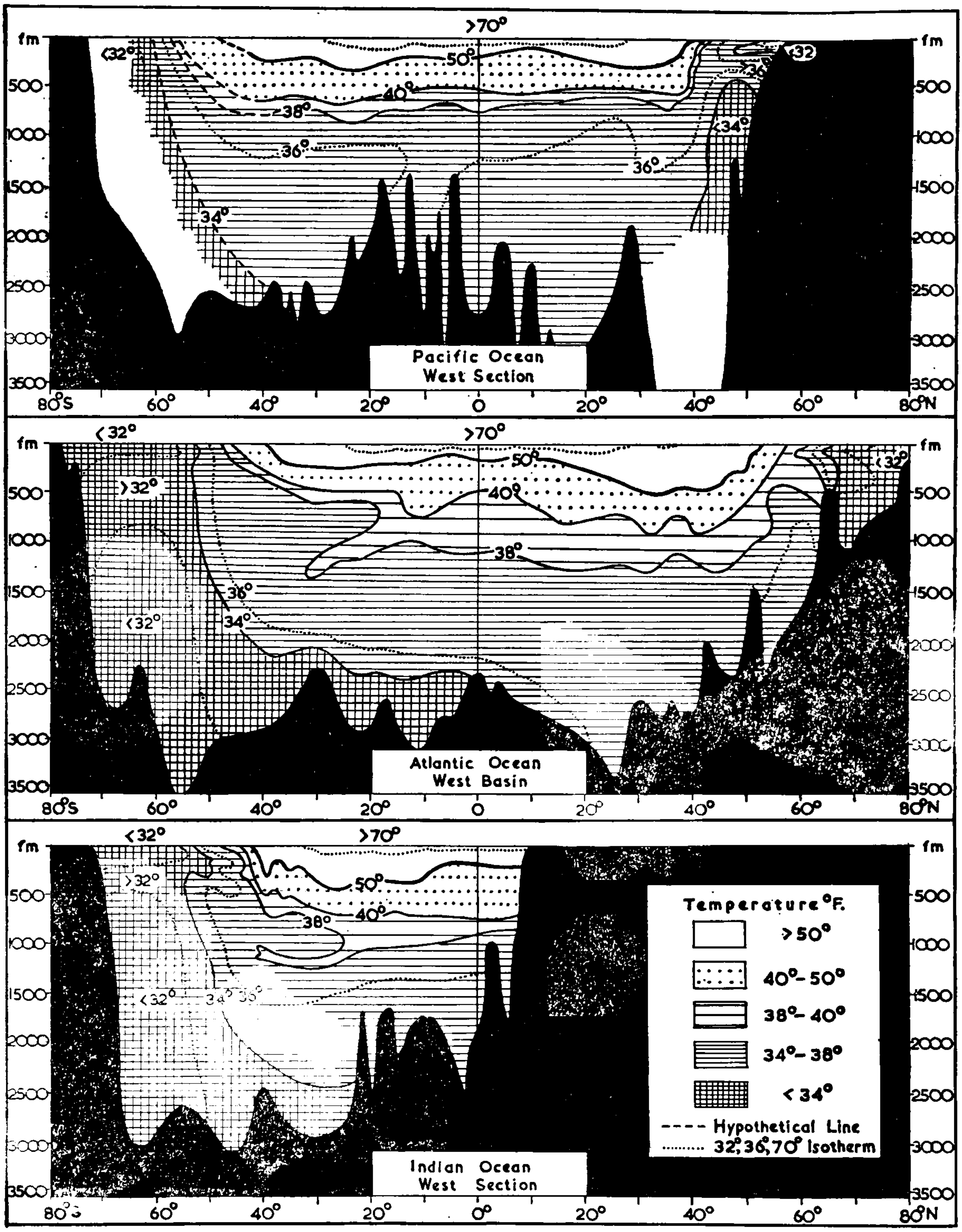


Fig. 25.1. Vertical distribution of temperature along sections (north to south) through the Pacific, Atlantic and Indian Oceans

Above this is a water layer of relatively high salinity flowing south, known as the North Atlantic Deep Current, which leaves the surface between Iceland, Greenland and Labrador. It receives a contribution from the Mediterranean outflow and continues southward to climb gradually over the underlying cold Antarctic water as this becomes progressively thicker. Above this layer is a third one which leaves the surface in the sub-Antarctic region and flows northward, forming the Antarctic Intermediate Current. This is detectable as far north as about lat. 20°N, and although cold it is less saline than the current coming from the north and so flows above it.

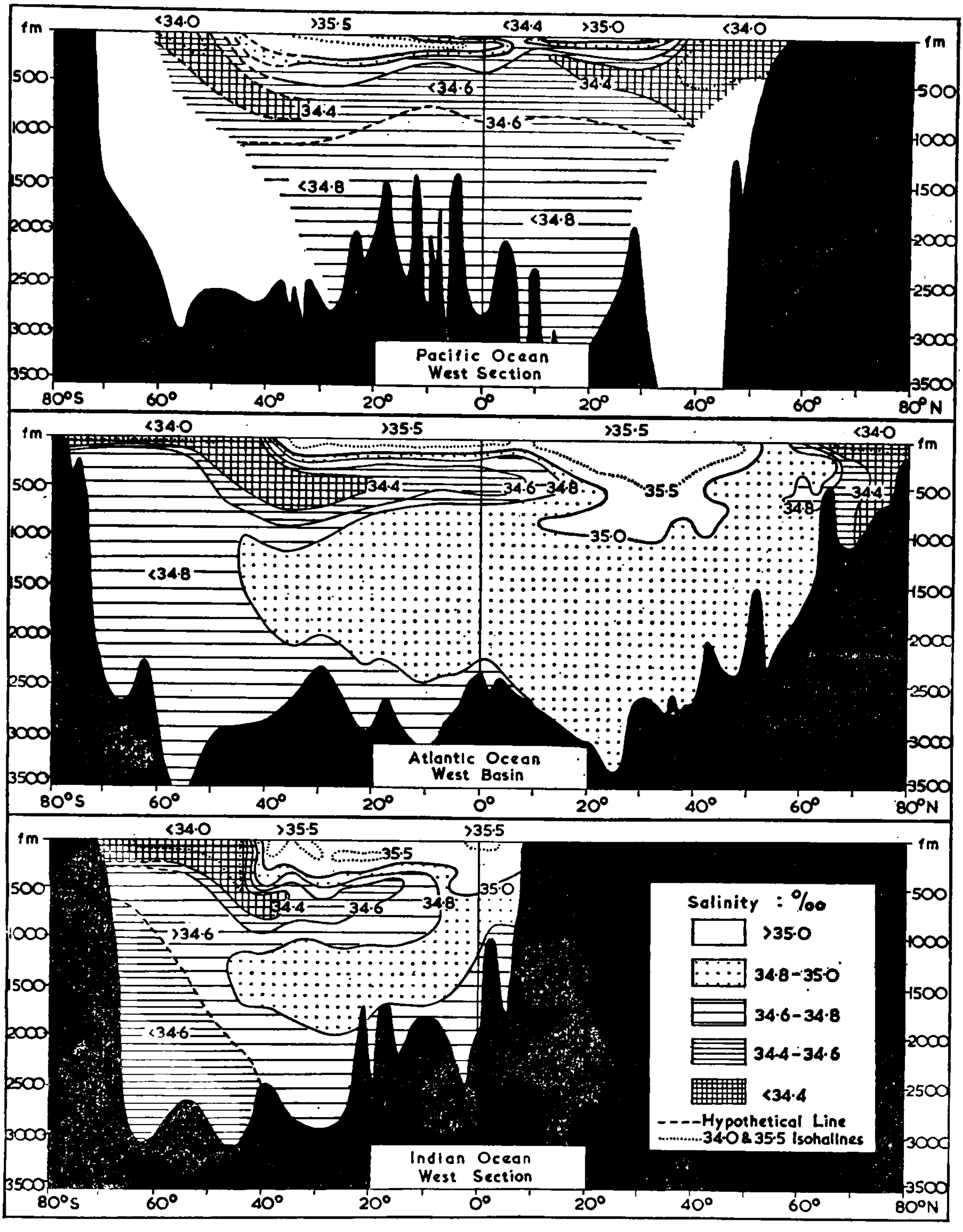


Fig. 25.2. Vertical distribution of salinity along sections (north to south) through the Pacific, Atlantic and Indian Oceans

Between 80 and 435 fm the temperature in the Atlantic is less at the equator than at lat. 30°N and 30°S, although the sea surface temperature is 9° higher at the equator. Between lat. 30°S and 10°N the bottom waters are less cold in the eastern basin than in the western—due to the Walvis Ridge. An examination of Figs. 25.1 and 25.2 will give some idea of the difference between the circulations in the depths of the Indian and Pacific Oceans and that of the Atlantic.

CHAPTER 26

SOME INTERRELATIONSHIPS BETWEEN OCEANOGRAPHY AND METEOROLOGY

Heat Capacity of Water

Thermally the oceans serve a dual function, to accumulate solar heat and to redistribute it in time and space. Their power to do this is derived mainly from two physical properties of water, its large heat capacity and its mobility. A given volume of water on cooling 1° could raise the temperature of 3,000 times as large a volume of air by the same amount. The mobility of the water in the vertical plane helps the oceans to accumulate solar heat, and its mobility in the horizontal enables the oceans to exert their thermo-regulating functions by means of warm and cold surface currents.

Because of the greater heat capacity of water, the extreme range of sea-surface temperature is only about one-third of that of the air at a few feet above the surface. This range is from 28·4°F in polar seas to about 104°F in regions such as the Red Sea, a difference of about 76°. The extreme range of air temperature is from about −90°F to 136°F, a difference of 226°.

H. Pettersson has compared the oceans to a kind of 'savings bank' for solar energy, in that they receive deposits during excessive insolation and pay them back in seasons of want. Some idea of the conditions which would prevail if there were no oceans can be got from comparing the climate of central Asia with that of the British Isles. The power of the oceans to distribute heat over the world's surface from their own store of heat (accumulated chiefly in the tropics) constitutes an extensive 'foreign exchange'.

Effect of Surface Currents

The warm North Atlantic Current—a continuation of the Gulf Stream—flowing north-eastward and northward to high latitudes on the east side of that ocean, explains the fact that the 32°F isotherm extends there beyond lat. 80°N, which is some 10° farther north than its average latitude for the whole hemisphere. The January air temperature off the Norwegian coast may be as much as 48°F higher than that for the same latitude averaged all round the earth. This is due to the warm current and the warmth directly transported by south-westerly winds. There seems no doubt that sea temperature has a fundamental influence in the incidence and the character of the seasons. It is possible that an unusually low surface-water temperature in large parts of the North Atlantic might be followed by reduced rainfall in a future season. Most of the rain over Europe is no doubt formed as the result of evaporation from the Atlantic, and the evaporation rate varies rapidly with sea surface temperature; a fall of 9° from 59°F to 50°F corresponds to a decrease in vapour pressure of about 30%.

Effect of the Atmosphere on the Sea

Perhaps the most obvious relationships between the two sciences concerns the winds, and the currents, waves and other characteristics of the seas which are ruled or influenced by the winds. For various reasons the average strength of wind over the ocean is higher than over land. The current-producing effects of wind are described in Chapter 16.

The effect of the wind in altering sea level has also been referred to in Chapter 16, in connection with current circulation. Local transitory winds affect the water level of the sea, sometimes with serious effect; winds from certain directions may cut this level in some rivers and harbours so severely as to hamper movements of shipping for a time. In rivers and estuaries such as the Elbe, sudden changes in water depth may be readily produced by certain wind conditions. This is especially noteworthy in the River Parana and in Rio de la Plata district. Nevertheless very high winds indeed are required to produce any appreciable change in water level off open coasts.

Thermal Effects of the Sea upon the Atmosphere

The main effects are:

(*a*) Direct heating or cooling of the lowest layers of the atmosphere due to contact with the sea surface, according to whether it is respectively warmer or cooler than the overlying air.

(*b*) The addition or removal of water vapour to or from the atmosphere by evaporation or condensation at the sea surface. This latent heat of evaporation is returned to the atmosphere at a higher level when condensation occurs at a later stage.

Energy Exchange between Sea and Atmosphere

In Chapter 1 it was stated that the mean temperature of the earth's surface varies little from year to year, implying that a balance must exist between the magnitude of the incoming and outgoing streams of radiation over the whole earth. An understanding of the factors involved in reaching this balance over the earth as a whole is needed in order to appreciate the problems of the energy exchange between the atmosphere and the sea.

In the first place all the solar radiation arrives as short-wave radiation and, according to the most recent figures (1954), 34% of the total amount is reflected back to space and lost. (This figure expresses the loss as an average over the whole surface of the earth and it is known as the earth's 'albedo'.) Of the remaining 66% about one-third (more exactly 19% of the incoming radiation) is absorbed by the gases, dust, water vapour and clouds of the atmosphere, while the other two-thirds reaches and heats the earth's surface, the tropics receiving a larger share than the polar regions. In between being absorbed by the atmosphere and the earth's surface and being radiated back to space, this fraction of the solar energy (66%) energises a meridional movement of air (i.e. N–S) at different levels in some parts of the atmosphere, which helps to reduce the excess of heat which would otherwise accumulate in equatorial regions, by transferring some of it to polar regions where a heat deficit prevails.

The return of this remainder of the solar energy (66%) to space is effected by means of a net outflow of long-wave radiation from the earth's surface and the atmosphere. This net outflow is made up as follows. First there is the amount of long-wave radiation which directly escapes. This amounts to 14%, expressed as a percentage of the total energy of the incident solar beam. It represents the difference between the rather larger total amount of long-wave radiation originally leaving the earth's surface, and the proportion of this amount (the 'back radiation') which is returned to the earth's surface due to scattering by dust and water vapour and re-radiation from the gases of the atmosphere and from clouds.

Condensation in the atmosphere of water vapour previously evaporated from oceans, lakes and rivers, results in a gain of energy to the atmosphere of 23% expressed as above, while convection currents (including the transfer of heat by turbulence) also afford a gain of 10%. Both these amounts are subsequently lost as long-wave radiation. The remainder of the total net outflow of long-wave radiation (66%) is accounted for by the return to space (as long-wave radiation) of the energy initially absorbed as short-wave radiation (19%) by the atmosphere. Of this 19% the gases, dust and water vapour originally absorbed 17% and the clouds 2%.

In considering the energy exchange between the atmosphere and the open ocean, certain differences in the physical properties of land and sea surfaces become important. (At the same time the properties of the atmosphere over the oceans are virtually the same as over land, except that the amount of pollution present is very much smaller.) The main properties wherein a sea surface differs from a land surface are: (*a*) in the amount of short-wave radiation which is reflected at the surface; (*b*) in the absorption of short-wave radiation beneath the surface; (*c*) in the amount of evaporation at the surface; (*d*) in the movement of surface water (drift), due to the stress of wind on a water surface; (*e*) in the transfer of heat by ocean currents.

When the sun is well above the horizon (elevation more than 40°) less than 5% of the incident solar radiation is reflected from a calm sea, and nearly all the energy penetrates the surface and is available to warm the water, nor does reflection become important until the sun's elevation is below 20°, *see* Table 22.1. The energy penetrating the surface is absorbed in a layer whose thickness is appreciable and depends upon the clearness of the water; away from land in sea water of average clearness over 90% of the incident energy is absorbed in the topmost 10 metres, in coastal waters the corresponding figure amounts to nearly 99%. Evaporation from the sea supplies much more latent heat to the atmosphere than can be supplied by evaporation from lakes, rivers and vegetation. In moderate and strong winds the rate of evaporation from the sea surface closely depends upon the wind speed at that locality and upon the magnitude of the difference between the vapour pressure of the air at the sea surface (assumed saturated) and the vapour pressure of the air as usually measured at the level of the ship's bridge. This vapour pressure difference becomes largest when cold and dry air lies over a relatively warm sea surface.

An important deduction from these facts, which has also been verified, is that the oceanic regions in the northern hemisphere which are subjected to the most intense evaporation lie a short distance to east of the American and Asian continents between lat. 30°N and 40°N and not, as might have been expected, in the tropics. The reason is because powerful outbreaks of very cold continental air frequently overrun the ocean off Japan and eastern North America in the region of the Kuro Shio and the Gulf Stream. In winter these are also the same regions where the greatest total energy exchange (i.e. the sum of the energy provided by evaporation and convection) is occurring, while in summer the regions where the total energy exchange is at a maximum are found in the tropical Atlantic and Pacific. The winter regions of greatest total energy exchange between sea and atmosphere coincide with the area where the largest proportion of temperate latitude depressions are initiated, while the summer regions broadly coincide with those where most tropical revolving storms begin their existence.

The movements of the surface water (drift) caused by the stress of the wind and the nature of ocean currents are described in Chapters 15 and 16. The influence on the climate exerted by the movements of the surface water layers down to the depths at which solar radiation penetrates seasonally is almost certainly an indirect one, which cannot be assessed as yet with the inadequate data available.

Difference between Air and Sea Temperatures

The average temperature of the surface sea water is slightly greater than that of the air. In the tropics the mean value of this excess is about 1·4°F. In middle latitudes there are large seasonal and regional differences. In the western North Atlantic and North Pacific the surface water in winter is as a rule warmer than the air by 6° to 8°F. In the eastern part of these oceans the difference is small. In spring and summer the sea surface is colder than the air in many regions, and on the Great Bank of Newfoundland the difference may amount to 2° or 3°F.

The air at the sea surface quickly adopts a temperature near that of the sea; thus the air temperature over the ocean will show on the average much the same distribution as the sea surface temperature, but comparatively slight differences between water and air are of great importance to the atmospheric processes, and are particularly significant for the purposes of weather forecasting. If the sea is warmer than the air instability occurs, whereby heat as well as moisture is readily propagated upwards into the atmosphere. The general slight excess of sea surface over air temperature therefore means that in most regions and during most of the year the sea is energising atmospheric processes by giving up heat and water vapour to the air. The bulk of the heat transfer, however, takes place in winter over rather well-defined areas, as described above.

The opposite state of affairs (air warmer than sea) means stability, and the cooling of the atmosphere will not be propagated anything like as quickly upwards as was the heating. Cooling of air by the sea, when the air is warmer, will therefore have little quantitative effect upon the atmosphere, compared with the heating when the sea is the warmer medium. As a result, the main influence of the sea upon climate is a levelling 'upwards'; the effect is to produce on the average a higher temperature (not only a smaller range of variation) and a greater precipitation.

Atmospheric conditions tend to be fairly uniform where there are only slight differences in temperature between air and sea, but locations of abrupt oceanographical changes are the scene of many and rapid meteorological changes. Such a region occurs along the northern border of the Gulf Stream, and small air displacements here may create pronounced stability or instability. Marked instability is often engendered in polar air invading the Gulf Stream area. Initially the air is much colder than the underlying sea surface. As a result the sea heats the lowest layers of air so rapidly that convection currents are unable to remove the heat to higher levels fast enough to prevent a steep lapse rate from developing near the surface. Another factor producing instability in the polar air is evaporation from the sea surface. The initial dryness of polar air favours this evaporation which is followed by condensation at quite low levels, and formation of instability showers as described in Chapter 4. In this case, and when more polar air than usual moves across the warm-water areas of the North

Atlantic to northern Europe for a few months or more, a surplus of rainfall and a reduced annual variation of air temperature may be expected over north-west Europe. Thus the ocean exerts an influence upon climate, although it is by no means a direct one. The distribution of isotherms of air and sea temperature over a region such as the North Atlantic shows that the influence of an ocean current (e.g. the Gulf Stream) on climate is considerable over a long period. However, when conditions over the same area are considered for a period of a few days, it is evident that the distribution of winds through this time is the factor mainly responsible for the heat exchange in the atmosphere between tropical and polar regions. Over periods of this length the influence of ocean currents on sea-surface temperature is probably slight; the main influence of the sea on the atmosphere over short periods is found in the powerful warming effect it has upon polar air masses moving towards lower latitudes, as described above.

An important difference between the characteristics of ocean and atmosphere, is that convection is brought about as a result of heating at the surface of the atmosphere but of cooling at the surface of the ocean. In both media, circulation is promoted and strengthened if a layer of higher temperature lies below one of lower temperature; if the reverse is the case, circulation is impeded and finally disappears.

Acknowledgements

Table 20.3 is adapted from *The Oceans*, by Sverdrup, Johnson and Fleming. Figs. 20.1 to 20.5 and 24.2 have been taken from the reports of the *Meteor* expedition. Fig. 21.1 is reproduced with permission of the Challenger Society from *Science of the Sea*. Figs. 24.3 and 24.4 came from an instructional memorandum issued by the French Hydrographic Office, and Figs. 25.1 and 25.2 have been adapted from two diagrams which appear in Professor A. Defant's paper, *Die Theorie der Meerströmungen und die Ozeanische Zirkulation*, 1929.

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