CHAPTER 3 GENERAL METEOROLOGY

3. GENERAL METEOROLOGY

3.1 Introduction

It is the purpose of this chapter to acquaint the reader with the environment in which typhoons form and move and to introduce such elementary meteorology as is required to understand the subsequent description of tropical cyclones and their behaviour. The nature of the cyclones of temperature latitudes and tornadoes are also discussed to point out some of the differences between these systems and tropical cyclones. Readers who find this chapter too demanding can omit it and should they encounter unknown terms later in the book they can consult the index where references will direct them to appropriate explanations. On the other hand, readers who are familiar with elementary meteorology as usually taught i.e. with a bias towards temperate latitude conditions, will find that the emphasis in this chapter is on tropical meteorology and tropical conditions, especially where these differ from conditions in higher latitudes; for these reasons, some of the sections may be of interest to them.

3.2 Pressure and density of the atmosphere

The atmosphere is a mechanical mixture of some 12 gasses but five of them - nitrogen, oxygen, argon, carbon dioxide and water vapour - make up 99.997% of the volume below 90 km. Water vapour comprises 4% of the volume (3% by weight) but is almost absent above 13 km. Ozone (a form of oxygen with 3 atoms in its molecule) constitutes a very small variable fraction of the atmosphere and is mainly concentrated between 15 and 35 km. There it constitutes a hazard to high flying aircraft - such as the Concorde - because it causes the decay of rubber components and can lead to ozone poisoning in passengers unless special precautions are adopted. On the other hand, the ozone layer is vi~~t~~ally important because it absorbs ulta-violet radiation with wavelengths between 0.2 and 0.31 μm, which would otherwise cause harmful effects to life on earth. Indeed, there has recently been concern that pollutants from high flying aircraft or from the widespread use of fluorocarbons might reduce the concentration of ozone to such a level as to give rise to increased skin cancer and other effects. Pollutants at these high levels are not scavenged by rainfall.

The pressure of the air at any level in the atmosphere is due to the weight of air above that level, from which it follows that the pressure will decrease with increasing height. The difference in pressure between two levels is due to the weight of air between them. This difference of pressure, expressed as a percentage, depends only on the temperature of the air; when air is warmer than normal it will be less dense than normal (because of termal expansion) and the pressure will therefore fall more slowly with height than in a similar column of colder air.

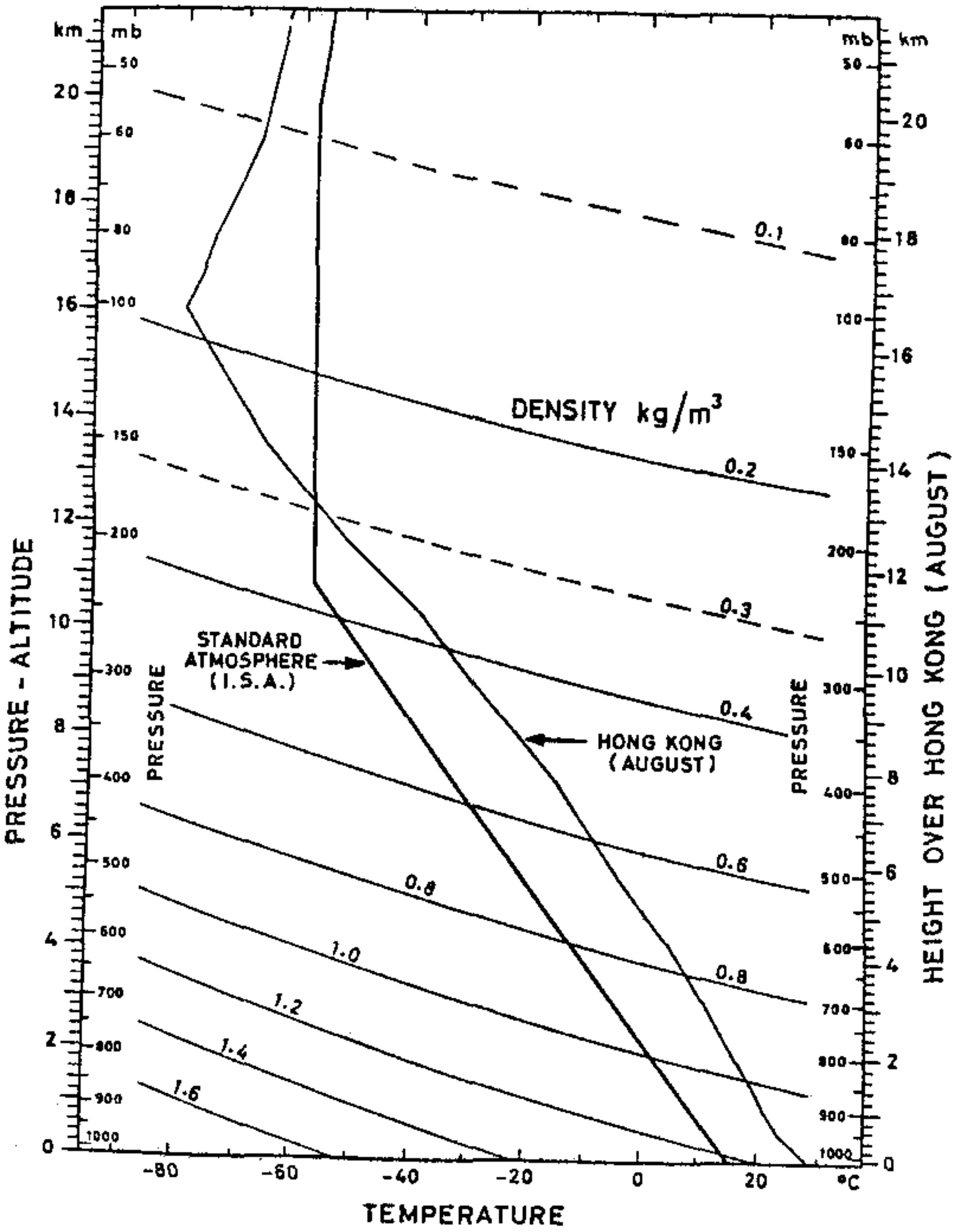

Fig. 3.1. The density of dry air at a given pressure and temperature is indicated on this diagram. The variation of temperature with altitude in the International Standard Atmosphere (I.S.A.) and at Hong Kong (August 1956/65) is also shown. The height (gpm) of the pressure levels in I.S.A. - pressure-altitudes - are shown on the scale on the left, those at Hong Kong in August are on the right. The I.S.A. is a dry atmosphere with pressure, temperature and density at sea level of 1013.250 mb, 15°C and 1.2250 kg/m^3 respectively and a lapse rate of temperature of 6.5°C/km up to 11 km then constant at -56.5°C to 20 km.

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Under average conditions the pressure decreases by a factor of approximately ten for every 16 km of height so that, the pressures at 16 km and 32 km are respectively 1/10 and 1/100 of that at the earth's surface. At 100 km it is only one millionth of that at the surface, whilst at 300 km the pressure is less than can be produced with a vacuum pump. It can be appreciated therefore, that there is no definite top to the atmosphere for the gases simply become more and more rarefied with height. It is believed that in interplanetary space there are usually between one million and one hundred million molecules per cubic metre; at heights at which the atmosphere has a similar concentration of molecules it will be indistinguishable from interplanetary space into which it merges. Because the air near the surface is compressed by the weight of that above, most of the mass of the atmosphere will be found at low levels. For example, in the tropics, approximately one half of the atmosphere by weight lies between the surface and 5.8 km and 90% lies below 16.6 km.

Air is a compressible gas and obeys Boyle's law. Therefore, if the temperature in the atmosphere is constant with height, the density will decrease as the pressure decreases. The density will fall by a factor of ten for each 16 km of height. In the real atmosphere, however, temperature changes with height and it is necessary to calculate the density from the prevailing pressure and temperature or use a diagram such as is illustrated in Fig. 3.1.

For some purposes use is made of a Standard Atmosphere (Fig. 3.1) in which the variation of temperature with height and the surface pressure are defined to be close to those prevailing in middle latitudes. In this atmosphere the surface pressure is 10.33 tonnes per square metre or, to use the meteorological unit, 1013.25 millibars (mb).* In 1964 the Italian physicist Torricelli found that the pressure of the atmosphere would support a column of mercury about 760 mm high in a vacuum tube.

* Footnote. In strict usage millibar should be abbreviated as mbar, but mb is in common usage and has the advantage of brevity, 1 mb = 100 Newtons (N) per square metre = 100 Pascals (Pa). A force of 1 N accelerates 1 Kg by 1 m/s^2. The hecto Pascals (hPa) is the W.M.O recommended S.I equivalent of the millibar and is slowly coming into general use.

Since this discovery, atmospheric pressure has been measured in terms of the length of the vertical column of mercury which it would support. The height of such a column also varies with its temperature (thermal expansion) and with its location on the earth because gravity is not everywhere the same. It is therefore necessary to correct the length of the column to a standard temperature ($0^{o}C$) and a standard value of gravity ($9.80665\ m/s^2$). The pressure can then be given as so many millimetres of mercury or the equivalent pressure in millibars. The standard atmospheric pressure of 1013.25 mb is equivalent to a corrected mercury column 760 mm high. Although atmospheric pressure is now measured in millibars in every country in the world a few, notably the U.S.A., still use inches of mercury domestically (and in aviation) and these units are found on some barometers. Barometers calibrated in inches are usually made to read correctly at a temperature other than $0^{o}C$. The standard pressure of 1013.25 mb is equivalent to a corrected mercury column 29.92 inches high. A conversion table for the different units of pressure will be found in Appendix 1.

The relationship between the pressure and height ~~at a level~~ in the atmosphere ~~and its height~~ is so significant that pressure (in millibars) is used by meteorologists to indicate levels; thus 1000 mb is close to sea level, 700 mb close to 3.1 km, 500 mb is close to 5.8 km and so on, the true height varying, of course, with prevailing pressure at sea level and the mean temperature of the air column. Near the surface of the earth the pressure falls by approximately 1 mb for every 10 m increase in height, at 20 km the pressure decreases more slowly at a rate of about 1 mb for 130 m increase in height.

~~From what has been said it should be clear that~~ Barometers can therefore be calibrated to indicate height above some datum instead of pressure; however, it is necessary to specify both the mean temperature of the air column between the datum and the instrument and the datum pressure for which the indication would be correct. It is usual to use the Standard Atmosphere for this purpose and the height above mean sea level (m.s.l.) indicated by an instrument so calibrated is defined as the "pressure - altitude". If the air below the altimeter has a mean temperature different from that in the Standard Atmosphere or if the m.s.l. pressure is not the specified 1013.25 mb then the indicated pressure-altitude will not be the same as the geometric height above m.s.l. Nevertheless, it is clear that aircraft flying on different pressure-altitudes cannot collide with one another. Pressure-altitude levels are therefore used on airways. Before descending for landing pilots adjust their

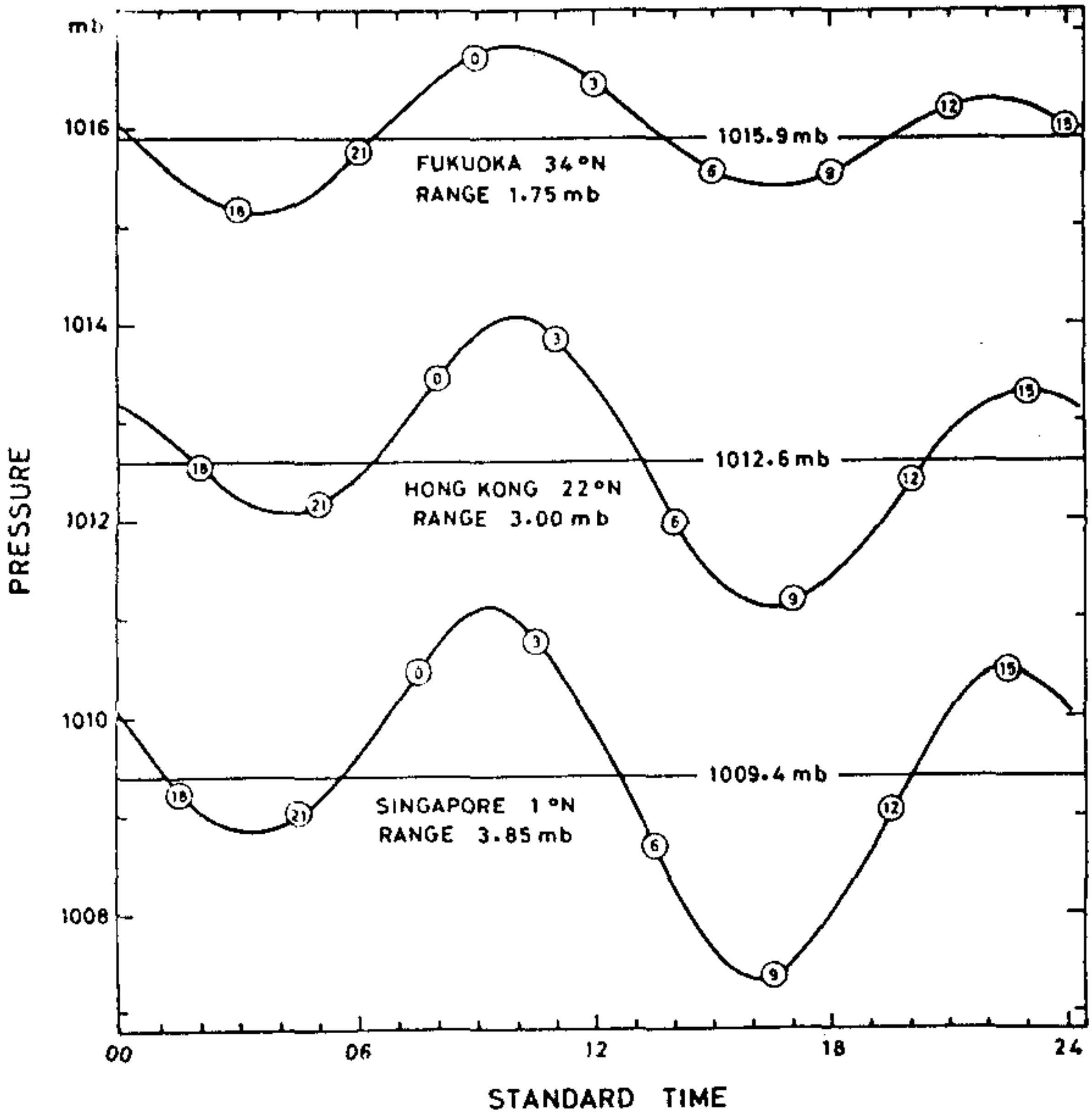


Fig. 3.2. The average variation of atmospheric pressure during the day at Fukuoka, Hong Kong and Singapore. G M T is shown in the circles on each curve.

altimeters for the difference between 1013.25 mb and the pressure on the airfield so that, on landing, the instrument will indicate either zero or the true height of the airfield above m.s.l. according to the setting used. There is, as yet, no international agreement on whether, at touch down, an aircraft altimeter should indicate "zero" or the height of the airfield above m.s.l. Errors arising from misinterpretation of the altimeter reading, changing to an incorrect setting, confusing inch and millibar settings, or using the wrong one of two altimeters with different settings have led to many aircraft accidents in the past and will, no doubt, cause more in the future until better procedures and instruments are found.

A weather map shows, among other information, the atmospheric pressure, corrected to m.s.l.at different places in the world. The correction to m.s.l. is necessary because stations are at different heights (and the pressure is less by approximately 1 mb for every 10 m of height). The pressure at a high station has therefore to be increased by an amount due to the imaginary column of air below it, making standard assumptions as to its density. Places on weather charts having the same pressure are joined by isobars which then delineate regions of high pressure (anticyclones) and low pressure (depressions or cyclones). By analogy with contours on a map a region of high pressure extending from an anticyclone is called a "ridge" and a region of low pressure extending from a depression or cyclone is called a "trough".

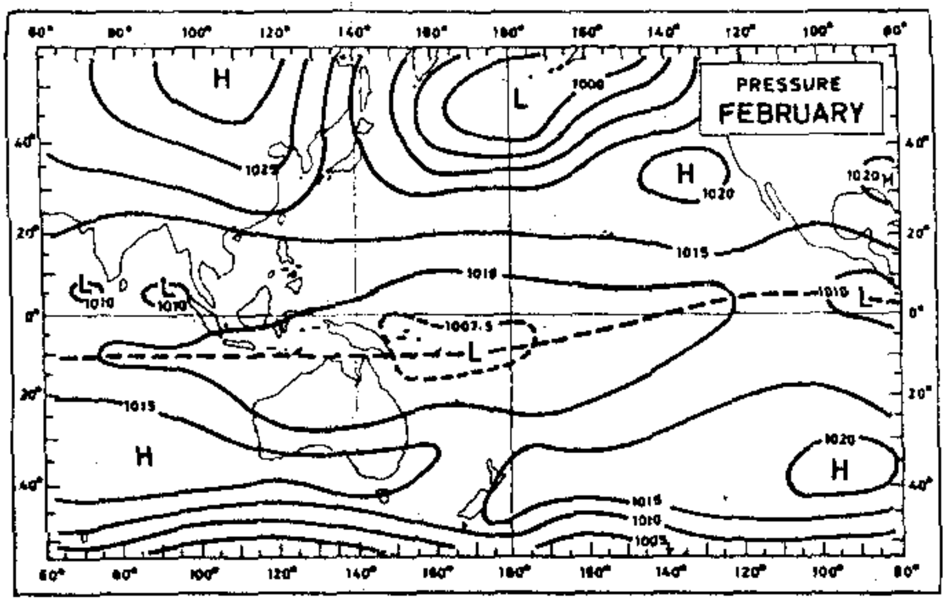
The pressure in the atmosphere has a diurnal variation being lowest at about 4 a.m. and 4 p.m. and highest at about 10 a.m. and 10 p.m. This variation (Fig. 3.2) is a monotonous feature of barograms in the tropics; it occurs also at higher latitudes but is of smaller amplitude and is usually masked by the changes due to passing cyclones and anticyclones. The diurnal variation of pressure is a tidal phenomenon caused by solar heating. It is believed that there would be only one wave each day were it not for the fact that the atmosphere is excited to resonate at its own natural period which, by coincidence, appears to be close to 12 hours. However, there is not yet complete agreement on the cause of the second wave (see e.g. Frost 1960). The range of the variation is not constant but changes with the time of year, cloudiness and other factors being greater in winter than in summer. In Hong Kong for example, the average diurnal variation ranges from 3.9 mb in January to 2.5 mb in July.

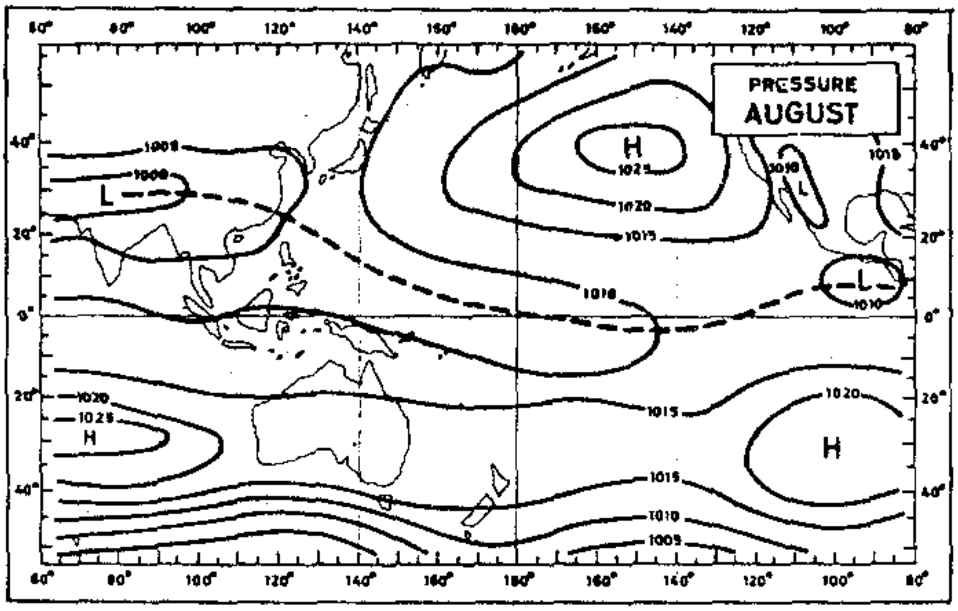
Changes from the normal diurnal pressure pattern in the tropics may indicate the nearby formation or approach of a tropical cyclone because they are regions of low pressure and cause the "mercury to fall". The usual signs are that the maximum of the barograph trace fails to develop or the minimum is much deeper than usual. For example, a "steady barometer" between 4 p.m. and 10 p.m. at Hong Kong should be read as a fall in pressure because a rise of about 2 mb would normally be expected during this period.

Reported mean sea level pressures have ranged from a low of 876 mb in typhoon June in November 1975 to a high of 1083.8 mb at Agata Lake ($66^{o}53'$N $93^{o}28'$E) on 31st December 1968 (Loewe 1969). Since the latter station is at a height of 263 m, the observed pressure has been increased by the weight of the imaginary column of air below the station. The air temperature was -46^{o}C at the time and even if the imaginary column is assumed to be 20^{o}C warmer the adjusted m.s.l. pressure still reaches 1080 mb.

Weather maps are drawn for different pressure levels because, as we have noted, pressure can be used as an index of height. Because such charts everywhere have the same pressure, isobars cannot be drawn. Fig.3.21c shows a chart for the 200 mb pressure level in February. The lines are contours and they show the height of the 200 mb pressure surface above m.s.l., just as contours on a map show the height of ground above m.s.l. Contours on upper-air charts perform the same role as isobars on surface charts; regions where the pressure surfaces are high correspond to regions of high pressure on a constant level map. This method of presenting information in the upper atmosphere was first used to meet the needs of aviation because aircraft cruise at pressure-altitudes and not at specific geometric heights. The custom is now universal and is used for both scientific and practical purposes. The levels for which upper-air charts are drawn are always expressed in millibars even in countries which still use inch units for surface pressures.

The potential energy per unit mass of a body due to the earth's gravitational field depends on its geometric height and the local value of gravity. If a body moves over the earth at a constant geometric height then its potential energy will change becuase of variations in the local value of gravity. In meteorology, it is useful to work with "geopotential heights" so that, in general, a body moving at such a height will not gain or loose potential energy. Geopotential heights are measured relative to m.s.l. and are expressed in "geopotential metres" (gpm) equal to ordinary geometric metres multiplied by g/9.8 where g is the local acceleration of gravity. At places where $g = 9.8\ m/s^2$, geopotential height will equal the geometric height. By international agreement, geopotential metres are now used universally for the measurement of height in the atmosphere and they are used in this book (e.g. in Fig. 3.1). In older texts the geodynamic metre (gdm) was sometimes used; this is related to the geopotential metre such that 1 gpm = 0.98 gdm.





b

Fig. 3.3. Monthly mean m.s.l. pressure over the Pacific and the Far East (a) in February, (b) in August. The axis of the equatorial pressure trough is shown as a dashed line. (After Crutcher and Davis 1969)

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3.~~3~~2.1 Surface pressure distributions

The lower atmosphere is essentially transparent to solar radiation which, by heating the earth's surface unevenly - depending on latitude, cloud cover, nature of the surface (water, desert, snow etc.) and other factors - causes the surface temperature to vary with location. The earth's surface, in turn, heats the overlying atmosphere by conduction, convection and radiation. Since the pressure of the air at any point is largely dependent on the temperature of the atmospheric column above, it follows that the surface pressure will also vary from place to place and with the seasons. The distribution of monthly mean pressure over the Pacific and the Far East in February and August is shown in Fig. 3.3. In the northern hemisphere winter, the North Pacific Ocean is warmer than the Asiatic mainland, the latter will then have relatively cold dense air over it, with higher surface pressures than over the ocean, Fig. 3.3a. In summer the situation is reversed and the hot mainland is then covered by the monsoon low whilst high pressure is found over the relatively cooler ocean Fig. 3.3b.

Four features of Fig. 3.3 are of the greatest relevance to the occurrence and evolution of typhoons, they are: 1) the Pacific anticyclone and its ridge - generally called the Pacific ridge - extending westward to China 2) the Siberian anticyclone 3) the equatorial trough, shown dashed, and 4) the Aleutian low pressure area called the "Aleutian low". The Siberian anticyclone and the Aleutian low are persistent and dominant features during the winter. The Pacific high is most intense in the summer. The equatorial trough covers the warmest parts of the earth's surface and so migrates north and south about two months behind the annual march of the sun. The average latitude of the Pacific ridge and the equatorial trough in different months are shown in Figs. 3.25 and 3. respectively.

3.8 <u>Temperature of the atmosphere</u>

It has long been known that the temperature of the air falls with height at a rate of about 6.5°C per km and it was considered that the temperature would continue to fall to the outer limit of the atmosphere. However, between 1899 and 1902 the Frenchman, Teisserence de Bort flew balloons bearing recording thermometers and so discovered that from about 10 km the temperature was almost constant or increased slightly with increasing height Fig. 3.1. Subsequently, it has been found that the level at which this change occurs varies from about 7 km in polar regions to 17 km in the tropics. It is known as the tropopause and separates the stratosphere above from the troposphere below. The troposphere, in which we live, is turbulent and contains vertical currents of air and associated clouds and weather. The stratosphere, however, is more stable and tends to remain layered, or stratified, so that rapid mixing does not occur there. Almost all weather (clouds, rain, snow etc) is therefore confined to the troposphere and aircraft flying in the stratosphere do so in relatively clear air. The troposphere should be visualised as a thin shell. Its depth of 10 km is only 1/1000th of the distance from the poles to the equator (10,000 km) and less than 1/600th of the radius of the earth (6370 km).

From information obtained from sounding rockets we now know that the temperature is warmer again above 20 km and is a maximum at about 45 to 50 km where it is almost as high as at the surface. Why is the temperature low at the tropopause and, indeed, why is there a tropopause at all? A small fraction of the incoming solar radiation is absorbed near 50 km at the top of the ozone layer

and causes high temperatures there, the remainder passes through the atmosphere without absorption until it reaches the surface of the earth where about half of it is absorbed and half reflected. There are thus two heat sources in the atmosphere, one at 50 km and one at the surface of the earth. Heat flows up from the earth by convection and long-wave (infra-red) radiation some of which is absorbed by water vapour in the lower atmosphere. The lower stratosphere contains very little water and is not heated by the outgoing

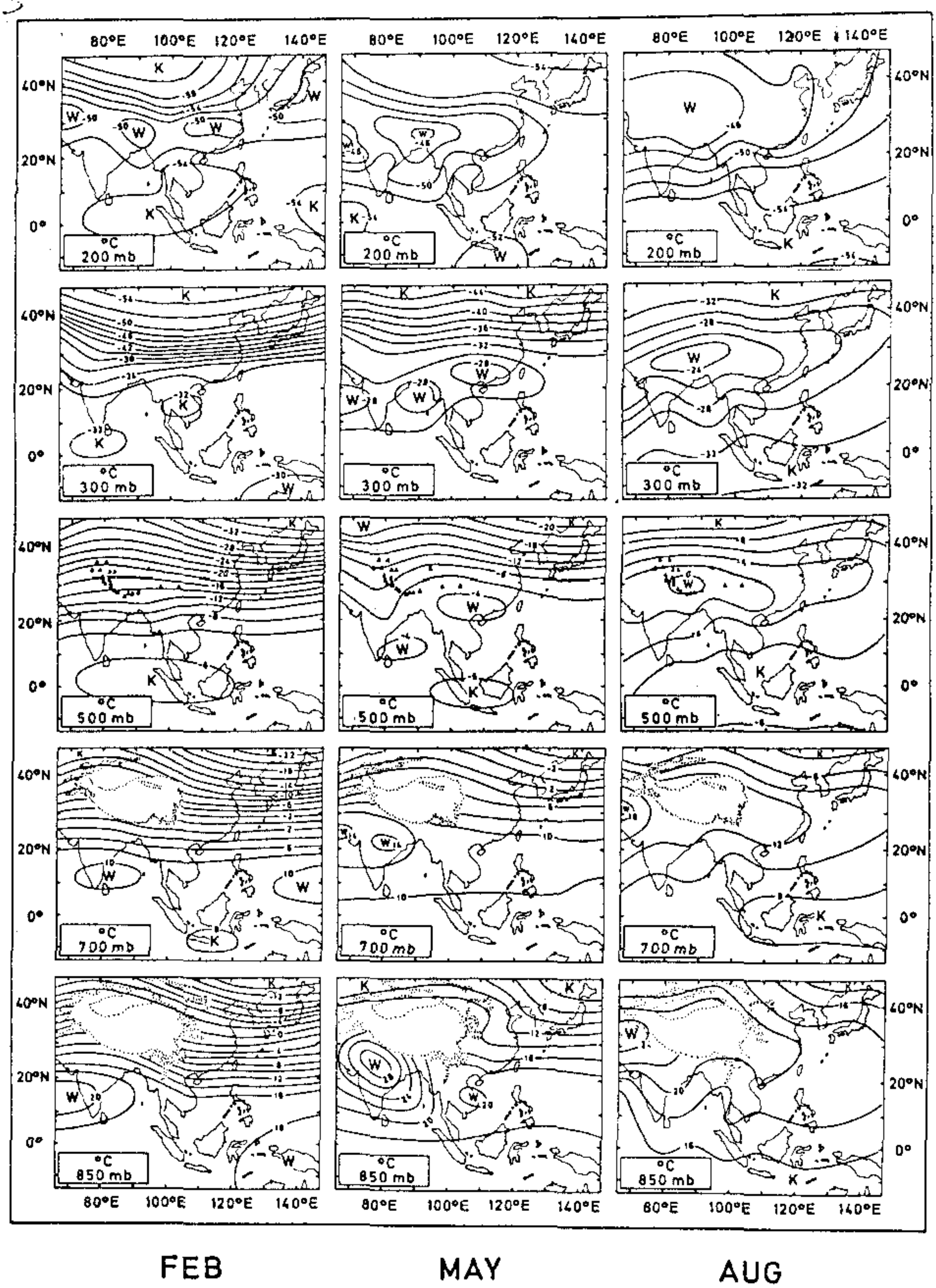


Fig. 3.4. The distribution of mean air temperature in °C at five levels during February, May and August. (After Chin and Lai 1974)

infra-red radiation and indeed, whatever heat it does absorb can soon be lost to space and to the earth by re-radiation. The tropopause is therefore a region where neither direct nor indirect radiation is absorbed or retained, it therefore remains as a layer of relatively cold air between the warmer layers near the surface of the earth and at the top of the ozone layer.

~~The ozone layer shields life on earth from much potentially damaging solar ultra-violet radiation which is largely obsorbed in the layer giving rise to the warming observed there. There is currently concern that the ozone layer might be depleted following chemical reactions with the exhaust emissions from airliners flying high in the stratosphere. The Concorde is not expected to fly high enough to cause any significant effect but the same cannot be said for supersonic aircraft designed to fly at higher altitudes. A more recently identified threat to the ozone layer arises from fluorocarbons - widely used in refrigeration systems and in spray propellants - which may diffuse up to the stratosphere, become dissociated by solar ultra-violet radiation and lead to the depletion of the ozone layer by chemical combination with oxygen.~~ out why?

3.3.1. Distribution of temperature

The distribution of sea surface temperature is discussed in section 14.1. Fig.3.4 shows the mean air temperature in the upper atmosphere at the main pressure levels. Warm regions are depicted by the letter W and cold regions by the letter K. There is a strong north to south gradient of temperature in the winter. In February, for example, there is a temperature difference of 34°C between the relatively warm air at 850 mb over India and the cold air over the Sea of Japan. In August this difference is only 10°C and the temperature over the Pacific is nearly uniform at 19°C. At 500 mb, solar heating during the summer months causes the temperature over the Tibetan massif to rise from the winter value of about 18°C to 0°C. This has a profound effect on the monsoons and the weather of South East Asia and causes the temperature to fall both poleward and equatorward of the massif.

3.4 Solar radiation

The sun emits electromagnetic radiation (Fig.3.4.1) with a wide range of wavelengths. However, the part of the spectrum that is of meteorological interest ranges from X-rays ($\lambda \approx 0.01\ \mu m$) to microwaves ($\lambda \approx 100$ mm). This region is shown in Fig. 3.4 2 along with the spectrum to be expected from an ideal black body+ at a temperature of 5780K. The departures from the idealized black-body spectrum are large in the X-ray and ultraviolet ($\lambda < 0.1\ \mu m$) regions and in the microwave bands. Radiation in these outer regions is highly variable and dependent on various solar disturbances. However, the total solar radiated energy is dominated by the central region of the spectrum between 0.3 and 10 μm and appears to be constant from day to day and year to year in so far as any changes are smaller than can be detected by current measuring techniques. Indeed, the flux of total solar radiation perpendicular to a plane at the top of the earth's atmosphere is known as the "solar constant" and has a value of 1367 W/m^2 with an uncertainty of less than $\pm$ 0.5%. This measurement was made by rocket-borne instruments in June 1976 and is considered the most accurate up to that time. Measurements of the solar constant made from within the troposphere are less accurate due to scattering and absorption effects (Fig. 3.   ).

Apart from electromagnetic-wave radiation the sun also emits corpuscular radiation, mostly hydrogen atoms and electrons, the quantities of which vary with the activity of solar disturbances. These particles take about a day to reach the earth whereas wave radiation takes only 8 minutes. The waves and particles cause changes in the upper atmosphere; X-rays ionise the upper atmosphere to form the E & F layers of the ionosphere which reflect certain radiowaves and permit radio communication around the earth. Solar flares are associated with increased corpuscular radiations which cause Auroral displays at the geomagnetic poles. Ultraviolet light produces the ozone layer.

+ An "ideal black body" is a hypothetical mass of material which absorbs all incident radiation (hence the term black) and emits the maximum possible energy (for its temperature) in all wavebands and all directions.

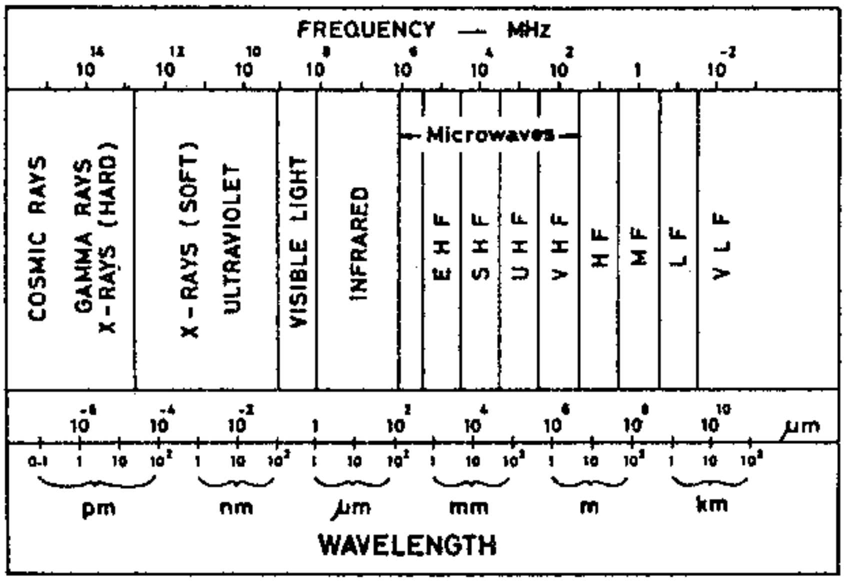


Fig. 3.4.1. The wavelengths and frequencies of electromagnetic waves from very short cosmic waves to very long radio waves (VLF - very low frequency).

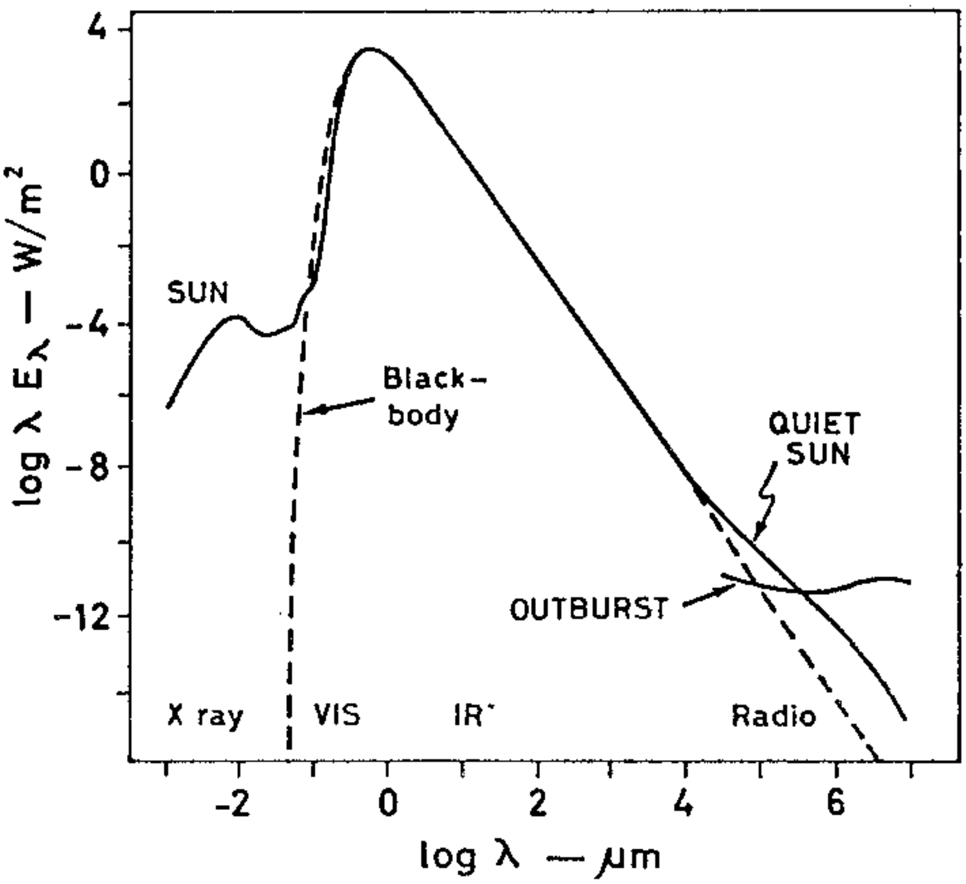


Fig. 3.4.2. The spectrum of the sun and a black bod at 5780K. (Adopted from Allen 1958)

Virtually all solar radiation with $\lambda<0.31$ µm is absorbed before it reaches the tropopause. Infrared radiation is absorbed predominantly in the troposphere where most of the water vapour resides. The atmosphere is transparent to visible light which covers the waveband from about 0.39 to 0.76 µm, coinciding with the peak of the solar emission. It is fortunate that there is such a "window" in the absorption spectrum of the atmosphere. The ground absorbs some of the solar radiation which is incident on it and radiates back to space with a maximum intensity in the infrared region between 10 and 15 µm. Atmospheric gases, particularly carbon dioxide and water vapour, strongly absorb these longer wavelengths (Fig. 3.4.3) so that much of the terrestial radiation is trapped and warms the troposphere. This warming effect is enhanced by the presence of cloud layers. High layers of cirrus (ice-crystal) clouds such as are found around and on top of tropical cyclones are most effective as a "blanket" because they not only block the escape of upcoming radiation but also loose a minimum amount of radiation to space because of their low temperature which may be -70°C or lower. The process by which the troposphere is transparent to incoming short waves but absorbs outgoing infrared waves is known in most text books as the "greenhouse effect". However, the effect itself is inappropriately named because, although it occurs in the atmosphere, it is not the prime mechanism at work in a greenhouse. The latter prevents the dissipation of the warmed inside air by wind or convection currents and so maintains it at a higher temperature than the outside air.

Neglecting the small amount of energy brought by corpuscular radiation and cosmic rays all the energy exchange between the earth and the universe is by electromagnetic radiation. To a close approximation the earth's atmosphere is neither gaining nor loosing energy in the long-term there is therefore a near balance between the incoming solar radiation and the outgoing terrestial radiation. This is illustrated in Fig. 3.4.4 where of 100 units of incident solar radiation 30 units are reflected back to space and 70 units re-radiated as long wave or infrared radiation. The earth intercepts solar radiation over an area equal to its cross section whereas the outgoing radiation is from the earth's total area of four times its cross section. On the basis of unit area of the earth's surface the incoming 100 units therefore correspond to the solar constant (1367 W/m^2) divided by four

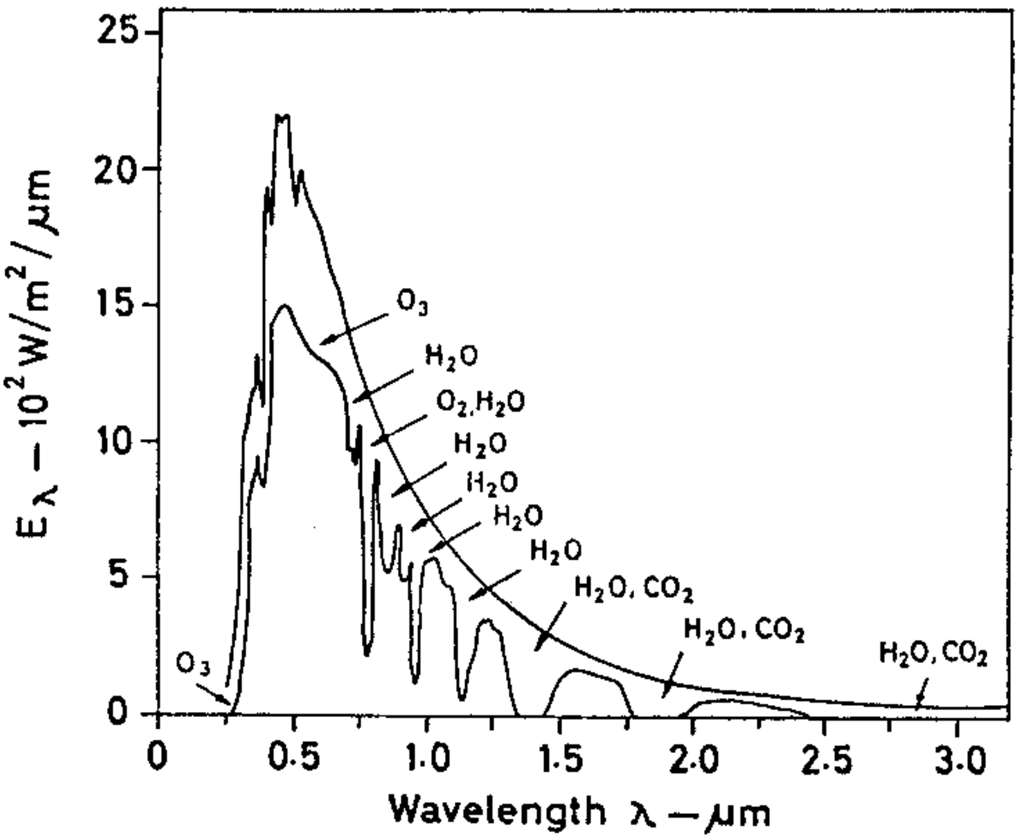


Fig. 3.4.3. Spectrum of solar radiation before and afer penetrating the atmosphere during average conditions and with the sun overhead. Areas of absorption by the main gaseous constituents are indicated.

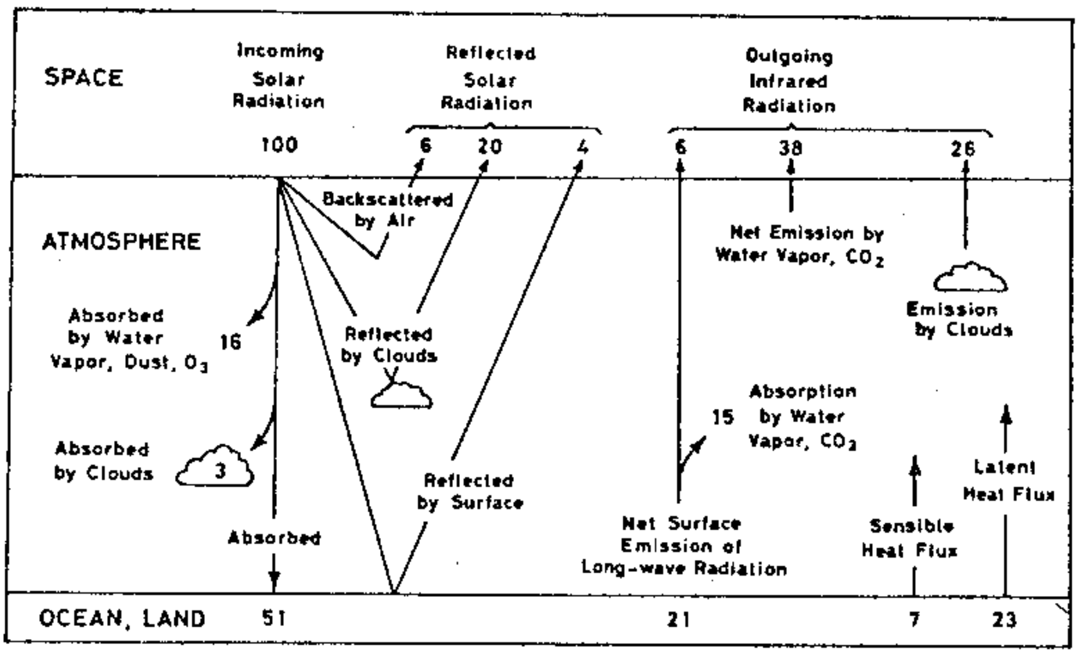


Fig. 3.4.4. The mean annual radiation and heat balance for the global earth-atmosphere system. The numbers represent percentages of the globally averaged solar radiation incident on the system. (Adopted from U.S. Nat. Academy of Sciences 1975).

or 342 W/m^2. The 51% of incoming energy absorbed by the earth's surface, is redistributed as a combination of infrared radiation and a flux of sensible and latent heats. The net infrared emission from the surface of 21 units represents the difference between a large upward flux and a slightly smaller downward flux from the atmosphere.

The Earth-atmosphere ratiation budget has been calculated using observations from the two-channel scanning radiometer abroad the NOAA operational satellites. The two channels observed reflected radiation from the earth in the visible range (0.5 to 0.7 μm)and emitted longwave radiation in the atmospheric intrared window (10.5 to 12.5 μm). Gruber (1977) presented results for the year from June 1974 to May 1975 and calculated that the fraction of incident radiation returned to space - the global albedo - was 32.3%. The outgoing longwave radiation was found to be 248 W/m^2. The net radiation Q of the Earth-atmosphere system is defined as

$$Q = I_o (1-A) - E \qquad (3.1)$$

where I_o = Incoming solar radiation

A = The Earth-atmosphere albedo

E = The outgoing thermally emitted radiation for the Earth-atmosphere system.

If the solar constant is taken as 1267 W/m^2 then ,for 1974/75, the net radiation received by the earth was

$$Q = \frac{1367}{4} (1 - 0.323) - 248 = 17 \, W/m^2$$

It is interesting that desert areas such as the Sahara have high albedo's of 40% or more and, being hot, emit copious longwave radiation thereby exhibiting net radiation deficits of 50 W/m^2 in latitude zones which are otherwise regions of radiation surplus. Over the oceans the high albedos associated with cloudiness in the equatorial trough remains north of the equator and varies little throughout the year. The variation of the net radiation by season and latitude is shown in Fig. 3.4.6.

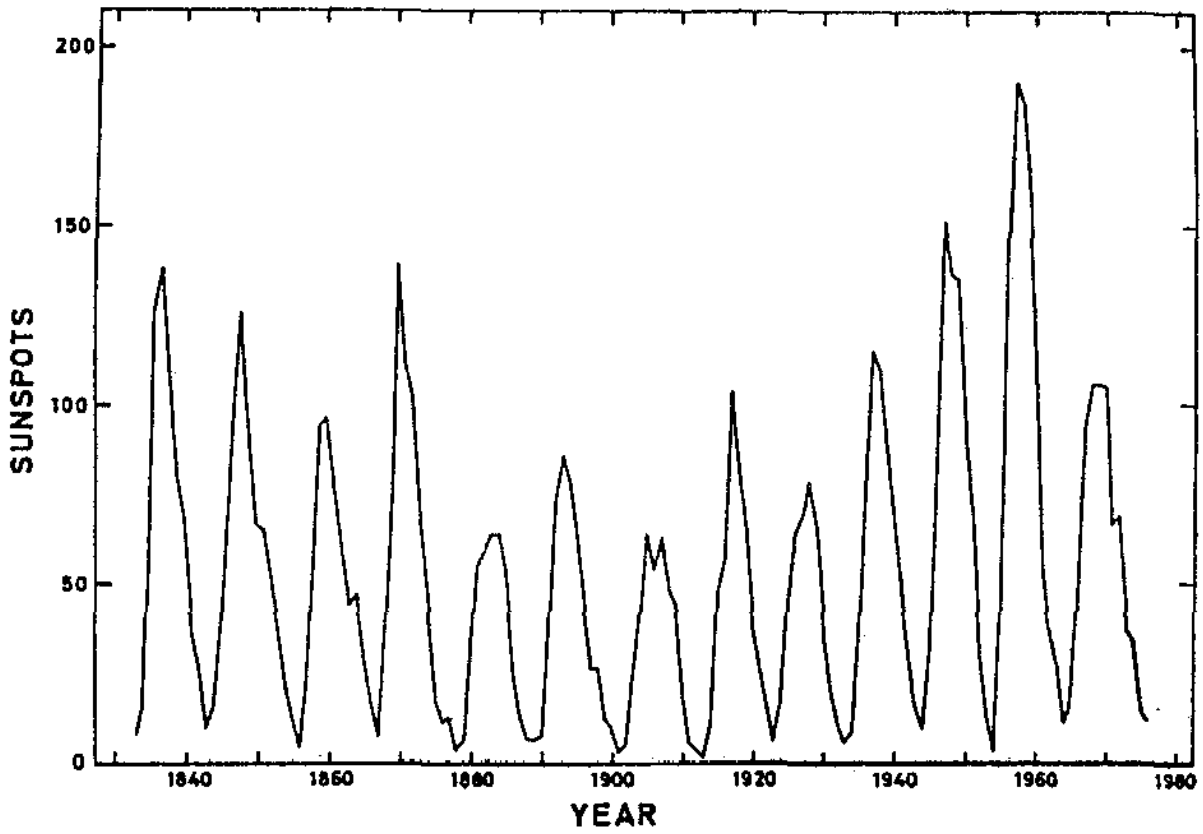


Fig. 3.4.5. Mean annual sunspot relative numbers. (Adapted from Bell 1977.)

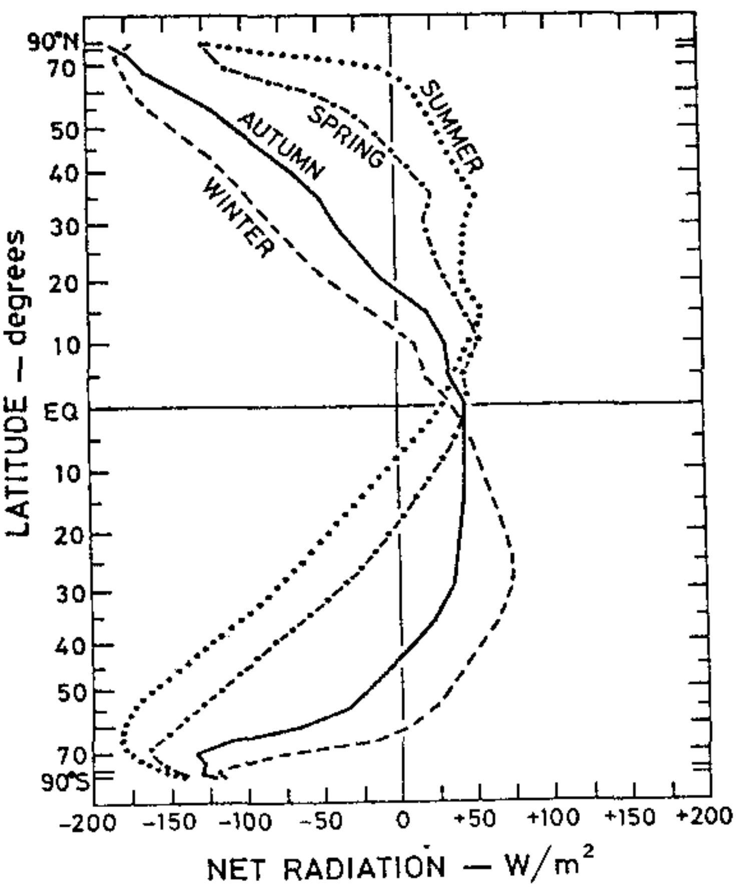


Fig. 3.4.6. Profiles of net radiation for the summer (June, July, August), autumn (September, October, November), winter (December, January, February) and spring (March, April, May) seasons. For the year from June 1974 to May 1975. (After Gruber 1977.)

3.4.1 Sunspots

Sunspots are dark, relatively cool areas which tend to appear in pairs in low latitudes on the sun. They vary in size, number and duration. A single spot will be between 10^3 and 200 x 10^3 km in diameter, there may be just a pair or as many as 50 grouped in two clusters and they can have a life cycle varying from hours to a few months. The longer-lived spots progress across the face of the sun and re-appear, two weeks later as the sun's 27.5 - day rotation brings them into view again. They are about 2000^{o}C cooler than the background temperature of about 6000^{o}C and are associated with intense magnetic fields. Pairs of spots are of opposite polarity and the associated magnetic loop within the sun is believed to interfere with the free convection of heat to the spots. The spottedness of the sun has been observed for over 200 years and is found to vary with an average periodicity of 11.2 years as illustrated in Fig. 3.4.5 although the duration of individual cycles has varied from 8 to 17 years. In one 11 year cycle pairs of spots have the same polarity to the east and those in the opposite hemisphere have a reversed orientation. In the next cycle the orientation of the polarities reversed. This reversal of polarity gives rise to a cycle with a 22 year periodicity which is sometimes called the hale cycle. There are therefore three periodicities associated with sunspots, 27.5 days, 11.2 years and 22.4 years. It is quite unknown to what the 11 and 22 year periodicity are due.

3.5 Vertical motion and clouds

3.5.1 Dry convection

When speaking of the fall of air temperature with height, meteorologists refer to the "lapse rate" and, as we have seen, an average "lapse rate" is about $6.5^{\circ}C/km$. However, on individual days and in different parts of the atmosphere the "lapse rate" will vary greatly from this average figure and, frequently, the temperature will rise with height over a considerable depth. In the latter case we say that there is an "inversion" of the lapse rate or, briefly, an "inversion".

We have already noted that solar radiation does not heat the lower atmosphere directly nor uniformly. Some surfaces present a more favourable aspect to sunlight than others; light coloured areas usually reflect heat whereas dark surfaces - such as tarred roads - usually absorb it; some materials do not conduct away absorbed heat and so become warmer than, say, a lake where the same amount of solar heat energy falling on an equal area will be shared by a considerable depth of water. Some parts of the ground will therefore become warmer than adjacent areas and, in consequence, some parts of the lower atmosphere will be warmer than their environment. This warmer air will be less dense than its surroundings and will be buoyant, in some ways like an air bubble in water. As the "bubble" of warm air ascends, and enters surroundings with successively lower pressure, it expands. When air expands without being supplied with heat it cools (conversely, compressing air causes its temperature to rise, as in a bicycle-tyre pump) but it will continue to rise, expand and cool until it arrives at a height at which its temperature equals that of its environment, it will then loose its buoyancy and come to rest. Decrease of temperature during expansion and increase during compression, without the addition or loss of heat, is called "adiabatic" expansion or compression. Dry air rising adiabatically will cool at the "dry adiabatic lapse rate" of $10^{\circ}C/km$ of ascent. Similarly, air subsiding adiabatically warms at the same rate.

If the environment air is "stably" stratified the "bubble" will become colder than its surroundings as soon as it is displaced and will return to its starting position; if the air is "unstably" strafified -

or "unstable" for short - then once it starts rising it will continue to do so spontaneously; all that is needed is a "trigger action" to start the ascent. A graph of height against temperature is shown in Fig.3.5a with a bubble of air, at temperature of 25°C, represented at A. The temperature in the free air - the "environment temperature" - measured by a balloon-borne thermometer is represented by the curve BD. The "bubble" A being warmer than its environment B accelerates upwards cooling at 10°C/km - line AC - until, at E, it attains the same temperature as its environment and looses its buoyancy. If a "bubble" were to rise past its equilibrium level it would become colder and denser than its environment and so sink - and warm - until it again had the same temperature as its surroundings. On fine days, without too much wind, birds can often be seen circling without flapping their wings as they use the rising currents, or thermals as they are called, to gain height. Sailplane flyers use these currents to fly great distances. The hot effluents from power station chimneys can be considered as visible thermals and when the wind is light they will be seen to rise almost vertically until they reach an equilibrium level, or inversion where they spread out horizontally. An inversion restricting the ascent of a thermal is illustrated in Fig. 3.5b; the bubble A can rise as high as in Fig. 3.5a only if it is sufficiently warmed at the surface to enable it to penetrate the lower inversion. If it is warmed to 26.5°C it will rise from F to G. Fig. 3.6 shows smoke trapped below a low level inversion which is penetrated by hotter effluents from a power station. This process by which heated air ascends by buoyancy is known as "convection". In nature there are many variations on the simple model of convection described here; in particular, thermals are weakened by mixing with the surrounding air as they ascend and they do not always reach the equilibrium level before being completely dissipated. If the hot source persists and if the wind is not too strong, succeeding thermals will rise in the warmed wake of the first until the equilibrim level is attained; ascent will thus take place by pulses.

It is not always necessary to have "hot spots" to cause air to rise by convection, a uniformly warm surface can also cause convection if its temperature is sufficiently high and there is some movement - usually eddies or turbulence caused by wind flow - to trigger the convection. In bright sunshine over uniform deserts the laps rate in the lower layer of the atmosphere may become even greater than the dry adiabatic (i.e. "superadiabatic")

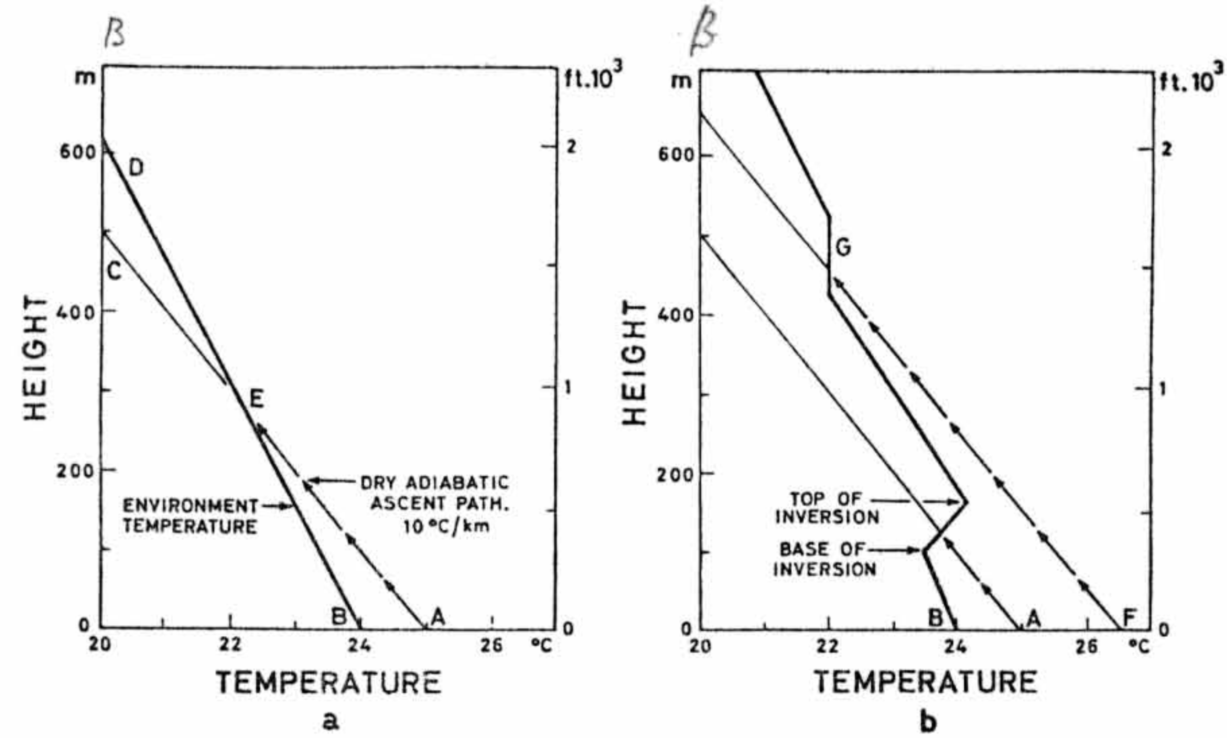


Fig. 3.5. The dry adiabatic ascent of air: (a) in an atmosphere in which the lapse rate is close to the average of 6.5°C/km and (b) in an atmosphere which has an inversion and a layer above, at G, which is isothermal. The parcels A and F, at 25° and 26.5°C respectively, are assumed to ascend without loss or gain of heat (adiabatically) cooling at 10°C/km.

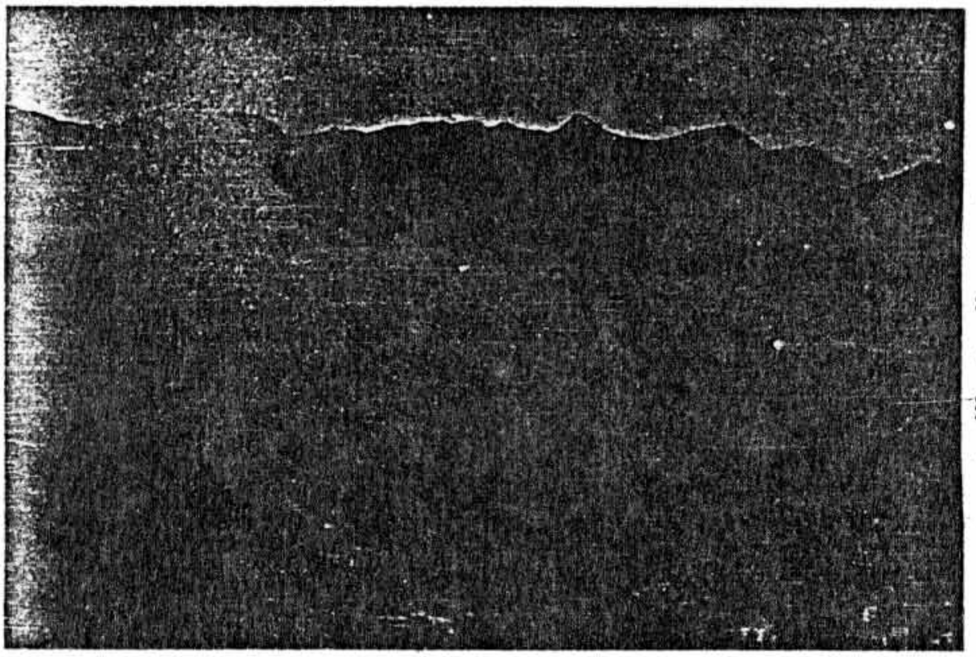


Fig. 3.6. Smoke and haze, over Hong Kong International Airport, trapped below an inversion which is at a level lower than the tops of the 550 m high hills seen in the background. Effluents from an electric power station were warm enough to penetrate the inversion.

but only small short-lived convection currents form and these are quickly eroded. The source area having lost its covering of hot air quickly becomes cooler than the surroundings and unable to support further convection. In these extremely unstable conditions once ascent is triggered by a hot source with large heat capacity or by mechanical turbulence, then rapid convection results. The drawing in of warmed air speeds up any rotation it may possess and a vortex is formed as in water going down a bath drain hole. The rotating column helps to reduce the erosion of the thermal which can proceed upwards for a few hundred meters, or more, carrying with it the dust raised by the vortex at the ground. This is called a "dust-devil". When there is a wind and there are no significant hot spots, rising currents tend to arrange themselves in patterns with near uniform spacing; the type of pattern being determined by the depth of the convection layer and the wind velocities therein. Convection also occurs when cold air flows over an already warmer surface such as a warm sea and in this case the air quickly picks up moisture and the convection currents become visible as convection clouds Fig. 3.9.

3.5.2 Water vapour

The amount of water vapour air can hold varies; we speak of dry and most air. Water vapour continually evaporates from the oceans and from other water surfaces into the atmosphere, however, there is a limit to the amount of moisture which air can hold at any given temperature and, when it holds this maximum amount it is said to be "saturated". At higher temperatures air can retain greater amounts of water vapour. This is illustrated in Table 3.1 which shows the number of grammes of water which one cubic metre of saturated air can retain in vapour form at different temperatures.

Table 3.1 Water Vapour Content of Saturated Air at Different temperatures

Temperature	^{o}C	-50	-40	-30	-20	-10	0	10	20	30	40
Saturation Water Content	g/m^3	0.1	0.2	0.5	1.1	2.4	4.8	9.4	17.3	30.3	51.1

The humidity of the atmosphere can be expressed in several ways; the measure used in Table 3.1 - grammes of water vapour per cubic metre of air - is known as the "absolute humidity". The "humidity mixing ratio" is the mass of water vapour in unit mass of dry air and is usually expressed as grammes of water vapour per kilogramme of dry air; the "specific humidity" is the mass of water vapour in unit mass of air plus water vapour, rather than in dry air alone and is usually little different from the mixing ratio.

The "relative humidity" is the ratio, expressed as a percentage, of the amount of water vapour held in the air to that which it could hold when saturated at the same pressure and temperature. If follows that if unsaturated air is cooled its relative humidity will rise until, at some temperature the air will be saturated - this is the "dew point temperature" or "dew point" at which the relative humidity will be 100%. Further cooling will cause water vapour to condense. In the humid tropics, the prudent hostess serves iced drinks in a glass enclosed in a jacket to absorb the copious quantities of water which condense on the cold glass where the air is cooled below its usual dew point of 25^{o} C or so.

In the troposphere the atmosphere always contains some water vapour which, if the air is cooled sufficiently, will condense on tiny particles called "condensation nuclei" to form a cloud of water drops. There are always large numbers of condensation nuclei in the atmosphere. When condensation takes place latent heat is released, and the amount released by a given mass of water is the same as was required to change it from liquid to vapour. Large amounts of heat are involved in these changes of state for example, about 2.5 million Joules* of heat energy are required to vapourise one kilogramme of water. The exact amount of heat required or released depends on the temperature at which the transformation takes place and varies from 2 500 kJ/kg at 0^{o}C to 2 257 kJ/kg at 100^{o}C.

* This is equivalent to 0.69 kilowatt hours or the heat emitted by a one kilowatt heater in 42 minutes. 1W = 1J/s so 1 kWh = 3.6MJ.

If the buoyant air in Fig. 3.5 is moist, and rises high enough to cool to its dew point so that condensation takes place, then latent heat will be released and the thermal will no longer cool at the dry adiabatic lapse rate but at some lesser rate, known as the "saturated adiabatic lapse rate" (Fig. 3.7). Near the ground in the tropics where the air is warm and can hold a lot of moisture the saturated adiabatic lapse rate will be about 4.5°C/km. Higher up, where the air is colder and cannot hold so much moisture, the saturated adiabatic lapse rate is greater and continues to increase with ascent until, at a height where the temperature is about -50°C, the dry and saturated adiabatic lapse rates are effectively equal at 10°C/km (Fig. 3.7b).

Rather than use graphs of height against temperature, meteorologists often use a more convenient diagram called a "temperature-entropy diagram" usually shortened to "Tephigram" or "TØ diagram" in which entropy is plotted against temperature Fig. 3.8. It is not necessary to discuss entropy here (see text books on thermodynamics) suffice it to say that when no heat enters or leaves a substance - as in adiabatic changes - its entropy does not change. Dry adiabats therefore appear on a TØ diagram as horizontal lines each one being identified by the temperature at which it crosses the 1000 mb isobar (isobars run from bottom left to top right); this temperature is referred to as the "potential temperature" and is expressed in Kelvins[1]. Apart from horizontal lines of constant potential temperature, vertical isotherms and sloping isobars the TØ diagram also contains curved saturated adiabats and lines of equal saturation mixing ratio (shown dashed).

The vertical distribution of pressure, temperature and humidity in the atmosphere as determined by sounding balloons can be plotted on a TØ diagram to determine - among other things - whether clouds will form by convection. An example of cloud formation similar to that of Fig. 3.7a is reproduced on a TØ diagram Fig. 3.8. From this figure it is possible to see how the water vapour content of an undiluted ascending thermal

1 The Kelvin (K) is the unit of thermodynamic temperature and is equal to the temperature in degrees Celsius plus 273.15 i.e. K = °C + 273.15. Temperature intervals expressed in Kelvins or degrees Celsius are identical.

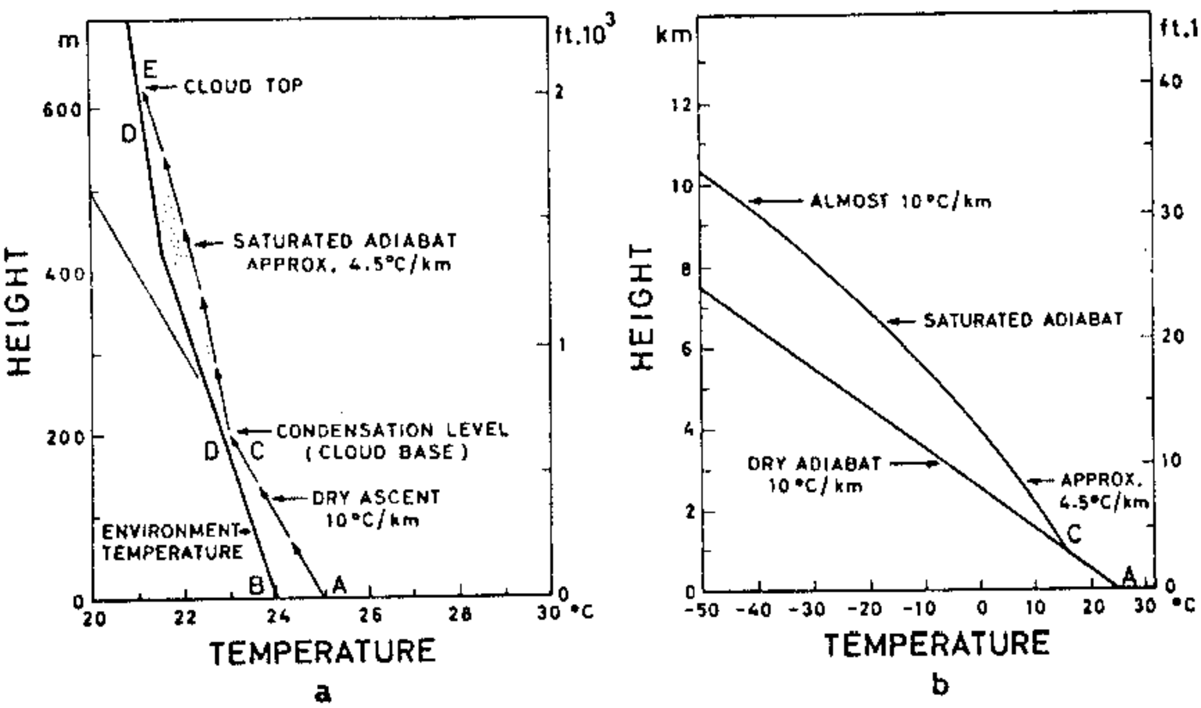


Fig. 3.7. Air at A which is warmer and less dense than its environment ascends by buoyancy, cooling at 10°C/km until at C it has cooled to its dew-point. Being warmer than its environment (D) it continues to ascend but release of latent heat slows the rate of cooling. At E it attains the temperature of its environment and ceases to rise. In (b) is shown a dry adiabat - 10°C/km - and a saturated adiabat, i.e. the path which air saturated at C would follow if always warmer than the environment. It will be noticed that the slope of the saturated adiabat gets closer and closer to that of the dry adiabat as the temperature decreases.

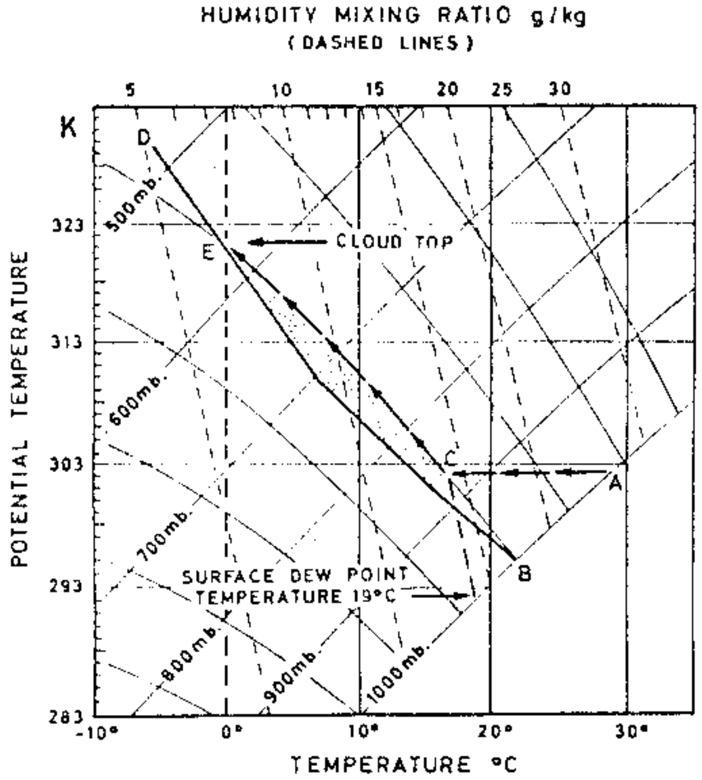


Fig. 3.8. Part of a "Tephigram" showing a plot of a temperature sounding (BED) and the adiabatic ascent of a parcel of air (ACE).

varies with height. The dew point temperature of the surface (1000 mb) air in the example is 19°C which corresponds to a humidity mixing ratio 14 g of water vapour per kilogram of dry air. Some parcel of surface air has its temperature raised to 29°C (A) and being warmer than its environment (B) it rises adiabatically along a horizontal line (dry adiabat) cooling at 10°C/km until at (C) it becomes saturated; thereafter it continues along a saturated adiabat (curved lines) condensing water vapour (with the release of latent heat reducing the rate of cooling with height) until at (E) it attains the same temperature (0°C) as its environment and looses buoyancy. At (E) each kilogram of air holds only 6.5 g of water vapour (read off at the top of diagram as indicated by a line through (E) parallel to the pecked lines) out of the original 14 g; the difference being condensed out to form the cloud. If the air could continue to rise along a saturated adiabat to 200 mb (12 km) its temperature would then be -58°C and water content 0.05 g/kg or less than ½% of the original amount. We can now proceed to discuss the clouds which are formed by this convective process.

3.5.3 Cumulus clouds

Clouds formed by the buoyant ascent of air are known as "convection clouds". In fine weather, when the distribution of temperature in the free atmosphere permits thermals to rise to the condensation level but not very much further, small convection clouds called "cumulus" form like tufts of cotton wool. They all have remarkably flat bases near the same level as in Fig. 3.9. This cloud-base level corresponds to that at which surface air becomes saturated after rising and cooling at 10°C/km; it is known as the "convection condensation level" (Fig. 3.6a). The height of the base of convection clouds can readily be calculated by multiplying the difference between the temperature and dew point of surface air by 121. Thus, if the air temperature is 30°C and the dew point 20°C the cloud base will be at 1 210 m. If the large volume of relatively warm effluents from power station chimneys (or other similar sources) contain sufficient amounts of water vapour and if they do not encounter either a low inversion to restrict their rise or vigorous turbulence to weaken them, then they too will form clouds. Two man-made cumulus clouds are shown in Fig.3.10.

If, with normal day-time heating, thermals are able to rise and form clouds then we say that the atmosphere is "unstable". If the instability - and distribution of moisture - is such that air can rise to between 3 km and 6 km approximately, then large clouds form; they are called "towering cumulus". The sharp edged, cauliflower appearance of the tops of these clouds Fig. 3.11 indicates that they are composed predominantly of water drops. Sometimes cloud towers will be seen to push upward vigorously and then quite quickly fade away, this occurs when they penetrate a dry layer of the atmosphere. As the air in a tower rises it mixes with, or "entrains" some of the surrounding drier air. Some cloud drops will evaporate into this relatively dry air causing cooling (evaporating water requires heat, it is the reverse of condensation). If the cooling is great enough the cloud will subside and evaporate. Entrainment reduces the buoyancy of all rising currents except those in the core of large cumulonimbus clouds.

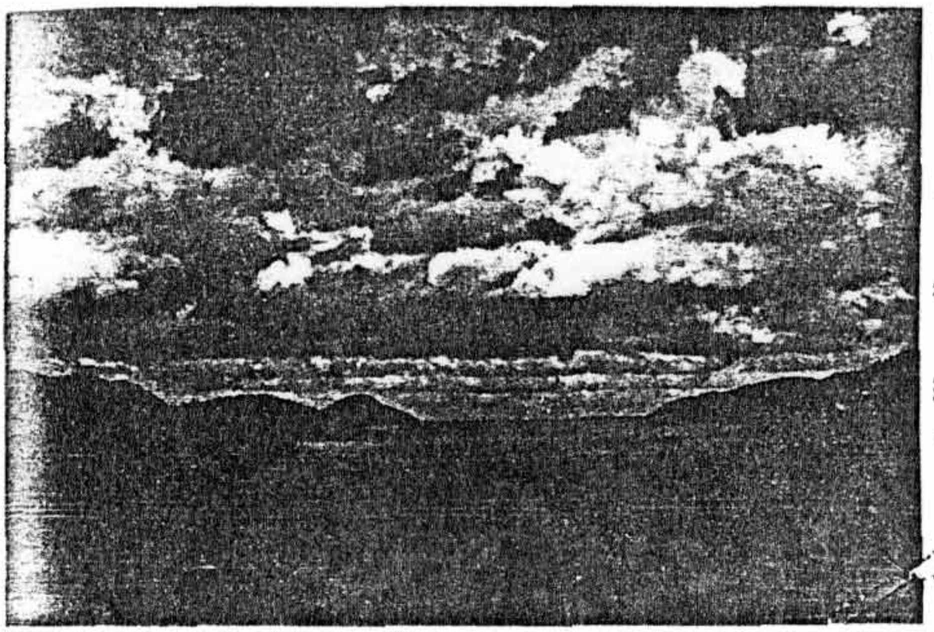


Fig. 3.9. Cumulus cloud streets over the sea. Note the uniform level of the cloud bases and the arrangement of the clouds in parallel lines running from left to right.

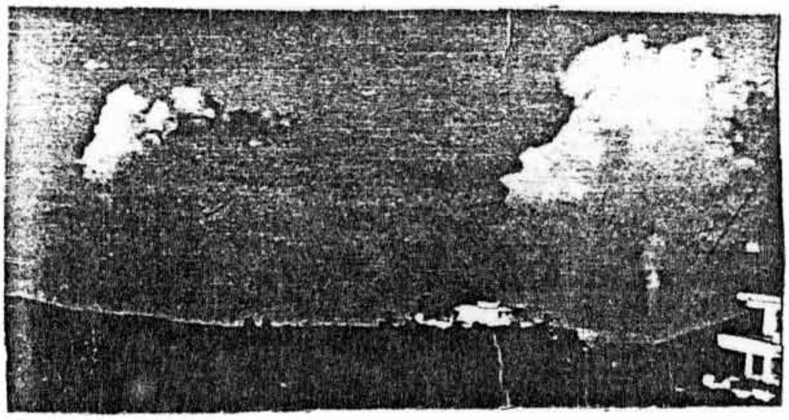


Fig. 3.10. Warm, moist effluents, from the electric power stations in Hong Kong's eastern harbour, are shown penetrating an inversion at about 600 m above sea level. Above the inversion the warm air continues to rise, as cumulus clouds, deriving additional buoyancy from the release of latent heat of condensation. Other pollutants, being relatively cool, remain trapped below the inversion. The top of the cumulus over the southern station (left) is being sheared off by an easterly wind at that level.

So far we have discussed only those convection clouds which arise from local "hot sources" but, as described earlier, convection also takes place over uniformly warm surfaces when, usually, the rising currents are nearly equally spaced or in long lines, known as "cloud streets" Fig.3.9. X
There is still much to learn about cloud patterns but cumulus streets usually lie along the direction of the mean wind in the convective-layer ~~wind shear (the difference between the velocity of the wind at the top and bottom of the layer)~~ and this, quite frequently, approximates the direction of the wind near cloud base. In typhoons, convection currents are organized in large spiral cloud bands and, in this case, the trigger to start the convection is the lifting caused by convergence as the air spirals in towards the typhoon centre.
The cloud bands may be hundreds of kilometres long and are usually spaced at intervals of 2 to 8km such that the ratio of the spacing to convective depth - called the "aspect ratio" is about three. The bands are most frequently associated with cold air advection over relatively warm surfaces. X

Sometimes air is "conditionally unstable" that is, it is stable unless it is saturated by the addition of water vapour or lifted to its condensation level and beyond until becoming warmer than its environment. This occurs when the lapse rate of temperature in the environment lies between the dry and saturated adiabatic lapse rates; lifting then cools the air at the usual 10°C/km until it becomes saturated and changes to the reduced saturated adiabatic lapse rate (approx 4.5°C/km) and reaches some level where it is warmer than the environment and can continue upwards by buoyancy forces. Lifting caused by the wind rising up over hills or masses of denser air (see Sect. 3.9) or in general convergence, can trigger the release of convection in conditionally unstable air.

3.5.4 Cumulonimbus

If the atmosphere is very unstable the buoyant currents may rise to the tropopause although they most frequently stop at some warm layer lower in the atmosphere. The clouds formed on such occasions extend well above the ~~freezing~~ 0°C level (about 5 km in the tropics Fig. 3.1) and their tops are composed of ice crystals and supercooled water drops. On reaching the tropopause, or other inversion, the vertical air currents spread horizontally to form an anvil shaped cloud (thunderhead) which, because it is composed of ice crystals, has a soft, fibrous, milky appearance Fig. 3.11 X
and 3.12. The edges are soft or diffuse because ice crystals, as they leave X
the main body of the cloud, evaporate less readily than do water drops.

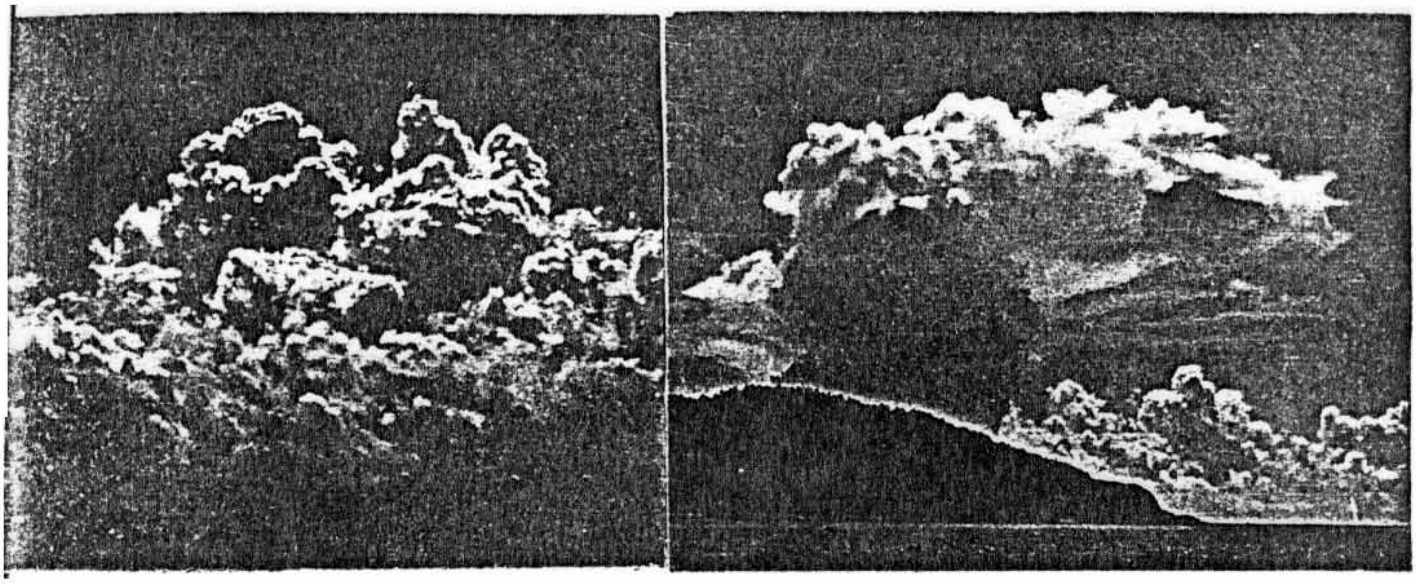


Fig. 3.11. Towering cumulus (left) and a developing cumulonimbus over the sea (right). The hard "cauliflower" tops of the towering cumulus clouds suggest that they are rising fast and are composed predominantly of water drops. Both clouds were near Hong Kong.

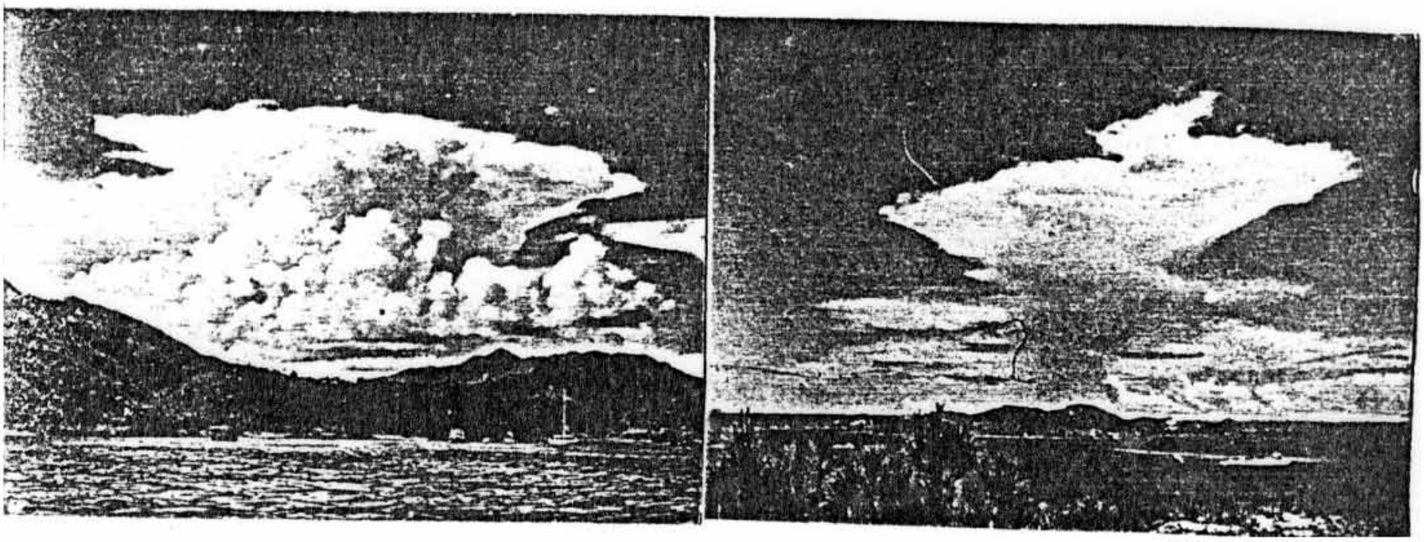


Fig. 3.12. Isolated cumulonimbus clouds with glaciated, spreading anvils. The cloud on left was over China mainland to the north of Hong Kong and that on the right was over the sea to the south of Hong Kong. Both clouds were about 50 km away.

These huge impressive clouds are called "cumulonimbus" and they are associated with heavy rain or hail, lightning, severe aircraft icing and extreme turbulence which, under unfavourable circumstances, has caused aircraft to disintegrate. ~~In~~ An extreme form of these clouds is the "supercell" (Sec. 15.7.1) in which ~~cases~~ the buoyant air ~~inside the clouds~~ can rise at velocities greater than ~~25~~ 50 m/s. Large hailstones form in association with these strong updrafts and theory indicates that stones with diameters up to 125 mm could be formed. The largest stone on record fell in Kansas U.S.A. on 3 September 1970 (Anon 1971) and weighed 766 g. The shape of the stone was irregular but was equivalent, in weight, to a sphere of water of diameter 114 mm. Hailstones with 'diameters' greater or equal to 100 mm fell in the Po Valley, Italy in three of the four years 1967-70 (Morgan 1973).

In the W.M.O. International Cloud Atlas, clouds are divided into genera which are further subdivided into species. The word "nimbus" after the name of a specific genera is used to indicate rain: thus comulonimbus is a raining cumulus. However, the genera cumulonimbus is specified as having a fibrous top - i.e. an ice crystal top - and/or thunder, hail and lightning. The term is not therefore applied to small cumulus or towering cumulus with hard cauliflower tops although rain might be falling from them.

The tropopause in temperate latitudes is usually at about 10 km whereas in the tropics it is frequently at 17 km and occasionally at 20 km. The tops of the largest clouds found in the tropics will therefore be much higher than those in temperate latitudes. The Dutch meteorologist van Bemmelen (1913) measured the height of cloud tops at Batavia (now Djakarta) in 1912, using cloud photographs and trigonometry and found some as high as 15 km. These measurements were not generally accepted in Europe until, in 1952, the first jet airlineer, the De Havilland "Comet" and early military jet aircraft began to encounter similar large clouds during flights in the tropics. Fortunately, supersonic aircraft like the Concorde, cruise sufficiently high above the tropopause to avoid cumulonimbus tops which would otherwise present a serious hazard to their safety. Airborne and ground-based radar which can "see" some snowflakes, hail and raindrops enable these aircraft to avoid cumulonimbus tops on the supersonic portion of the climb.

The highest cumulonimbus clouds form most frequently over land in the humid tropics and measurements of the height of their tops have been made using aircraft, radar and photographic techniques in several such areas including Malaya, Bengal and around Darwin, Australia. There is no doubt that cloud tops in these regions sometimes extend to at least 21 km and occasionally reach as high as 6 km above the tropopause. Over Bengal, Cornford and Spavins (1973) found tops at 20 km with one cloud top at 17 km still growing upwards at 8.6 m/s. Over Malaya, Hill and Lewis (1974) used radar to measure cumulonimbus tops up to 21.4 km. They found that the highest tops occurred most frequently over the lee side of the land mass - the highest clouds being found on opposite sides of the peninsula in the southwest and northeast monsoons.

Turbulence has been found 3 km above the top of these large clouds and 25 - 30 km around them. Frost (1954), using instrumented aircraft, found the average diameter of the large cumulonimbus clouds over Malaya and Sumatra to be 10 km. The tops of towering cumulus do not change to ice clouds at the 0°C ~~freezing~~ level but at some greater altitude where the temperature is well below -10°C. Frost found the change from water cloud to ice cloud to take place most frequently between 9 and 10 km where the temperature was between -30°C and -35°C.

Tropical thunderstorms yield very high rates of rainfall over periods up ~~of one~~ to three hours and often lead to deep and rapid flooding. ~~Instantaneous~~ Rates of rainfall over 3 s ~~(15 s mean)~~ of more than 550 mm/h ~~357 mm/h~~, have been measured in thunderstorms at Hong Kong with hourly totals in excess of 157 ~~100~~ mm. Once in 50 years, on average, thunderstorms in the China Sea area yield average rates of rainfall over 15 s, 20 min and 1 h of about 430, 200 and 120 mm/h respectively (Bell and Chin 1968). A radar presentation of a large cumulonimbus cloud containing heavy rain is shown in Fig. 10.

3.5.5. Stratus

We have noted that air does not rise by buoyancy alone; stable air can be forced to ascend as, for example, when it converges towards a cyclone centre, rises over mountains or over a mass of denser air. When such large scale ascent of stable air takes place the vertical velocities are of the order of 0.01 m/s or about one hundredth of those commonly found in convection clouds. The resulting cloud forms in sheets spread horizontally over large areas but usually having relatively little vertical development. These layered or strati-form,clouds are called "stratus", "altostratus" and "cirrostratus" and are usually found in the tropics below 2 km, between 2 and 8 km and between 7 and 20 km respectively. Rain can fall from thick layers of stratus type clouds in which case they are called "nimbostratus". Cirrus clouds are always composed of ice crystals. Cirrostratus sometimes forms by the spreading out of many cumulonimbus anvils. For example, in summer, the South China coast sometimes lies under a veil of cirrostratus blown southward from the anvils of many giant cumulonimbus clouds which form over the relatively hot interior of China. Towering cumulus, cumulonimbus, altostratus and cirrostratus all play important roles in tropical cyclones.

3.6. Theories of Rain Formation

The smallest drops to fall out of clouds have radii of about 100 microns[+] (0.1 mm); smaller drops fall very slowly and usually evaporate a short distance below the cloud. Drops with radii between 100 and 500 microns fall as drizzle, those greater than 500 microns (0.5 mm) are, by convention, called rain. The largest raindrops are about 6 mm in diameter (radius 3 mm); falling drops greater than this tend to break into smaller drops (Lenard 1904). Blanchard (1950) showed that drops as large as 8 mm can exist for periods of minutes in a uniform non-turbulent flow. Subsequent experiments and calculations showed that distilled water drops with diameters as large as 9.1 mm can persist in non-turbulent environments (Ryan 1976). However, although raindrops with diameters as great as 8 mm have been observed in very heavy thunderstorm rain in Japan and Korea (Shiotsuki 1976) this is exceptional, drops greater than 6 mm are only rarely found in a natural environment. A typical cloud droplet has a radius of 10 microns (0.01 mm) and simple arithmetic will show that it takes one million such droplets to make a typical raindrop or radius 1 000 microns (1 mm). Theories of rain formation have to explain how such remarkable growth can be achieved.

+ 1 micron = 1 millionth of a metre or 1 thousandth of a millimetre.

In the atmosphere there is an abundance of small particles which act as nuclei upon which water vapour can condense to form cloud drops. Many of these particles are hygroscopic i.e. water molecules condense on them before a relative humidity of 100% is attained. As a result of the abundance of such condensation nuclei rain does not form by condensation alone because big drops - having a relatively low concentration of hygroscopic material - will grow only slowly whilst condensation takes place preferentially on fresh nuclei with a greater concentration of hygroscopic material. Therefore, when the relative humidity rises, water vapour tends to condense on new nuclei to form new drops rather than increase the size of existing drops. Clouds therefore tend to have large numbers of small droplets all of more or less the same size - radii 5 to 30 microns approximately - and all falling relative to the rising air at about the same speed of from 10 to 100 mm/sec. Furthermore, the growth of drops by condensation alone is a slow process. A cloud drop of radius 1 micron having formed on a typical salt nucleus of mass 10^{-12} g* will, in a saturated atmosphere, take over four hours to grow to a drop with radius 30 microns which is still only a fraction of the size of the smallest raindrop (500 microns).

3.6.1 The Wegener-Bergeron process

In 1935 the Swedish meteorologist Tor Bergeron, in a classical paper, proposed a theory that rain is formed when ice crystals and super-cooled water drops co-exist in a cloud. He received strong support from the German scientist Findeisen but it was originally expounded in 1911 by Wegener and so is sometimes referred to as the "Wegener-Bergeron" process (Mason 1971). To understand this process it is necessary to appreciate that cloud drops can be carried upwards past the level at which the temperature is 0°C without freezing, we noted this fact in section 3.5.4 when discussing cumulonimbus clouds. Drops can remain liquid at temperatures as low as -40°C but freeze at lower temperatures or if they encounter active freezing nuclei. Ice nuclei are relatively rare in the atmosphere, particularly those which are active at temperatures of a few degrees below zero. In most supercooled water clouds therefore, there will also be a variable number of ice crystals.

* 10^{-12} corresponds to the fraction 1/1 000 000 000 000.

Water molecules, always in motion, continually escape - evaporate - from water drops but if the drop is in equilibrium with the surrounding water vapour there will be just as many molecules caught - condensed - on its surface as are lost, and the drop will not change in size. However, an ice crystal in this same environment will not be in equilibrium, it will gain more molecules than it looses because they are bound more securely in ice crystals than in water drops at the same temperature. The ice crystals therefore grow at the expense of the neighbouring supercooled water drops. In technical terms, the ice crystals grow because their equilibrium vapour pressure is less than that for water at the same temperature; at -10°C for example, the difference in vapour pressure between the two states is 0.27 mb. The rapidly growing ice crystals soon acquire substantial fall velocities and collide with smaller ice crystals so growing more by "coagulation" to become snowflakes. When they fall below the freezing level the snowflakes begin to melt to form waterdrops which, being more compact than snowflakes, fall faster until they finally reach the ground as rain. The process can be seen on a radar screen (Fig. 3.13) because as soon as snowflakes begin to melt they can be "seen" better than either snowflakes or raindrops (sect. 10.2); the "melting band" therefore appears on radar scopes as a "bright band" just below the freezing level. Below the melting band the faster falling, compact raindrops accelerate away leaving fewer drops to be "seen" in a given volume.

The coalescence process

Marchand (1903) reported a case of rain falling from a warm cloud i.e., a cloud which is everywhere warmer than 0°C. He observed the cloud to be 1 700 m thick below the observatory on the Pic du Midi in the foothills of the Pyrenees. Meteorologists in the tropics also observed that rain fell frequently from warm clouds. However, at that time, it was not possible to measure the temperature at the top of particular clouds nor to be certain that no large raindrops or snowflakes were falling onto the cloud from above. For this reason, and also because Bergeron and Findeisen had stated categorically that all moderate and heavy rain was formed by the ice process, most meteorologists treated these reports as suspect. However, Heywood (1940) obtained at Hong Kong the first proof (Mason 1971) that warm clouds could yield significant rain. In August 1939,

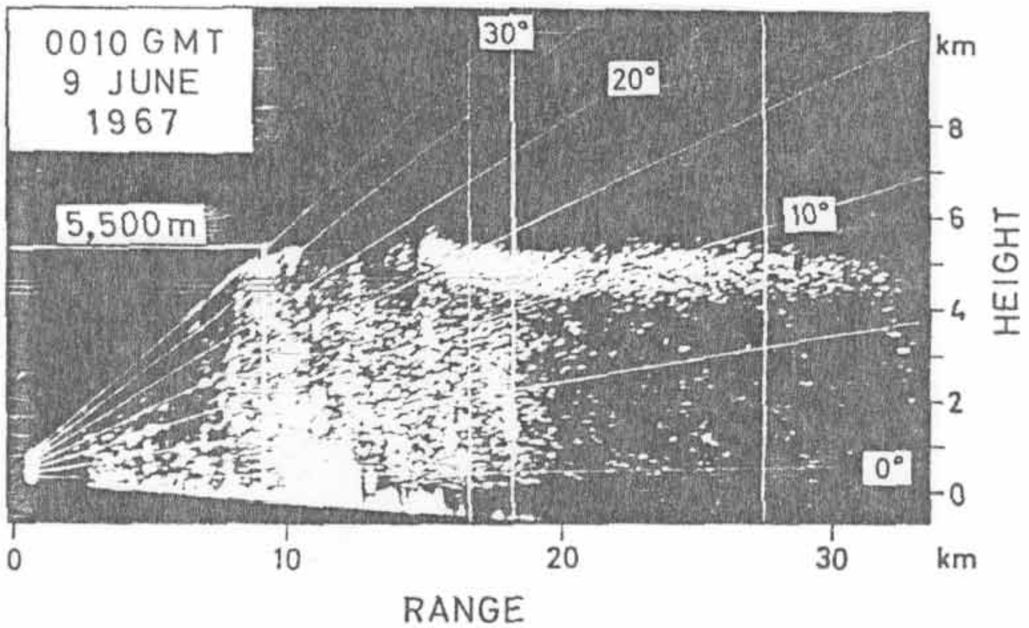


Fig. 3.13. A photograph of the RHI display of the Hong Kong Plessey Type 43S radar (wavelength 100 mm). The melting band is shown below the freezing level at 5.5 km. The radar was looking in the direction 218° and its sensitivity was decreased by a factor 1/300 (25 db). The approximate rate of rainfall beyond 20 km is 18 mm/hr and between 10 and 20 km it is 100 mm/hr in places. The heavier rain patches could still be seen when the radar sensitivity was reduced to 1/10 000 (40 db) of the normal value. On this display a given patch of rain would appear equally bright at all distances from the radar.

using an instrumented aircraft, he measured the temperature of the top of a cloud to be 13°C whilst it was raining at the rate of 3.2 mm/h over the Royal Observatory. As with Bemmelen's cloud top reports, many years passed before meteorologists generally accepted the observations. Pilots flying in the tropical Pacific in World War II reported seeing many rain-producing clouds warmer than 0°C and these observations helped to obtain general recognition, by the mid-1950s, that in the maritime tropics moderate rain frequently fell from "warm clouds" with tops between 2 km and 3 km (Fig. 3.14 and 3.15). Using instrumented aircraft Frost (1954) and many others have since confirmed the earlier observations. Ohtake (1963) has estimated that 53% of Hong Kong's rain falls from warm clouds and that, in some islands in the tropical Pacific, the percentage is even greater. How does this rain form?

Rain can form in a warm cloud by the process of "coalescence" if the range of cloud drop sizes is large. A drop larger than its neighbours falls faster and will encounter many smaller drops in its path some of which will coalesce with the larger drop which thereby grows larger than before, falls faster than before, collects more cloud drops than before, grows larger....... and so on; some drops reach a size at which they are unstable and break up under the stresses caused by their motion, thus providing a fresh supply of "large" drops. If the cloud is deep enough, contains updrafts of about 1 m/s or more and persists long enough then this chain reaction will soon form rain. The warmer the air the more water vapour will be condensed and the greater will be the water-content of the cloud i.e. more or bigger cloud drops per unit volume. For this reason a warm tropical shower cloud need not be as deep as a temperate latitude shower cloud. Frequently the top of a temperature latitude cloud would reach the freezing level before growing tall enough to produce rain by coalescence. These are the reasons why warm cloud rain is more frequently observed in the tropics.

Only a small fraction of the volume of a cloud contains water. The liquid water content of a good tropical cumulus cloud will usually be less than 1 g/m^3 and in colder clouds it may be only one tenth of this value. One gramme of water occupies one millionth of a cubic metre therefore, less than one millionth of the cloud volume contains liquid water!

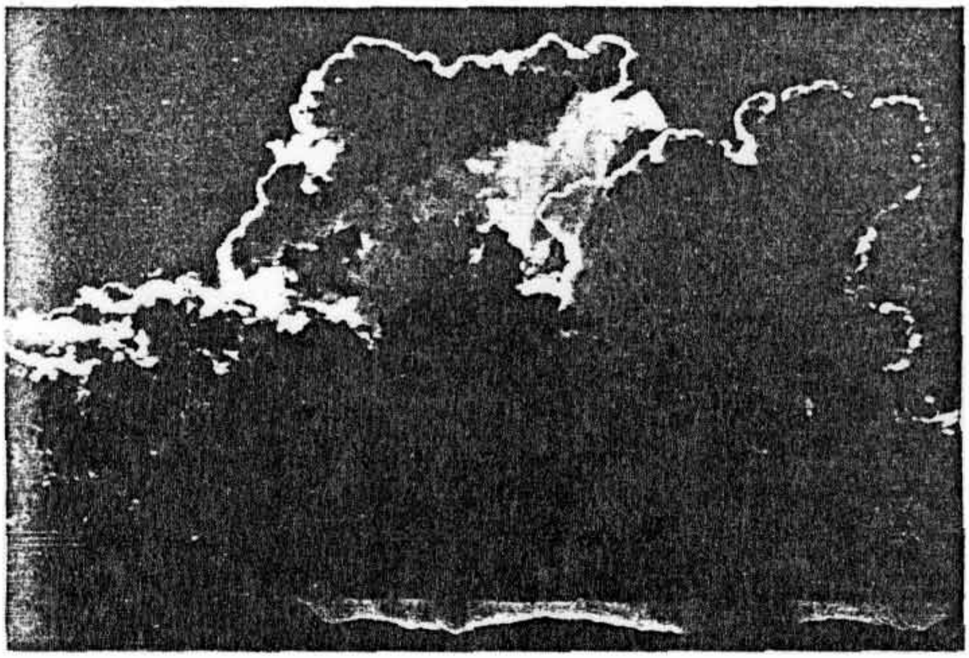


Fig. 3.14. Large tropical cumulus clouds producing showers by the coalescence process. The cloud on the left is beginning to decay, the small cloud drops evaporate first leaving the larger ones and raindrops which fall to produce a tenuous, streaky cloud.

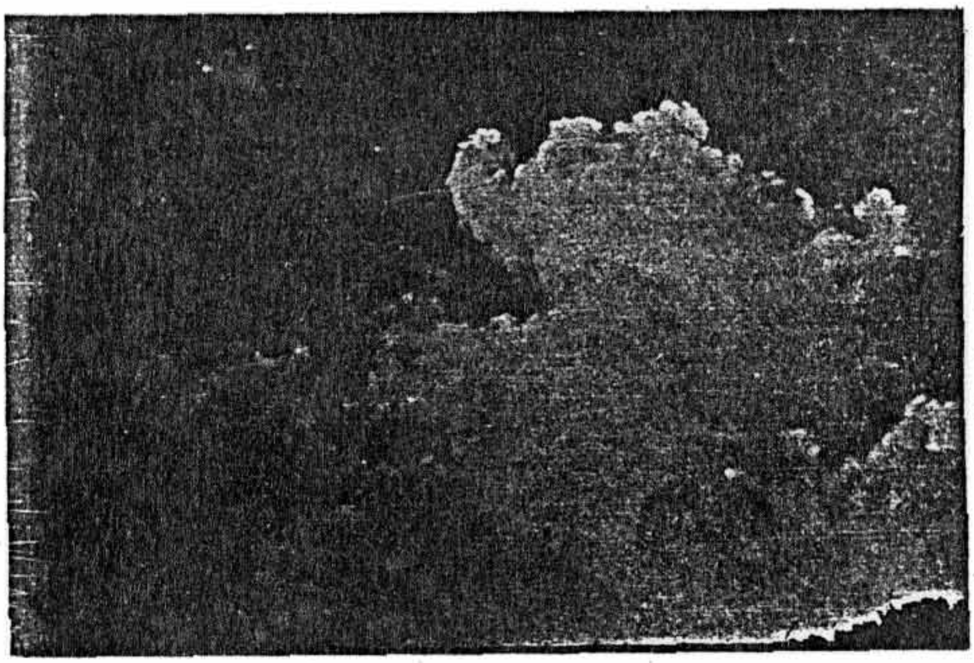


Fig. 3.15. A similar cloud to that in Fig. 3.12 showing spectral colours on the left due to the refraction of sunlight by falling raindrops, as in rainbows.

How does one drop become larger than its neighbours to start the coalescence process? For many years scientists have been measuring the size and numbers of both condensation nuclei and cloud drops. Woodcock (1950) found that air near coasts contains some "giant nuclei" with radii from 2 to 20 microns; these nuclei are salt particles that have been formed over the oceans by the evaporation of sea spray and the bursting of bubbles. He found that there were from 1 000 to 100 000 such nuclei per cubic metre of air. This concentration is similar to that of rain drops but is much less than the 300 million cloud drops found in the same volume in cumulus clouds. Ludlam (1951), showed that under certain conditions giant nuclei would grow to have radii of 30 to 50 microns and that they could play the role of the larger - than - average drop necessary for the coalescence process to begin. Jonas and Mason (1974) have shown how successive thermals rising with entrainment into one cloud could produce a bi-modal distribution of drop sizes such as is required to initiate the coalescence process. However, there is still much to learn about rainfall mechanisms. To summarise, we may note that in middle latitudes although rain may occasionally fall from warm clouds in the vast majority of cases it is intitiated by the Wegener-Bergeron process. Both the coalescence process and the Wegener-Bergeron process are important for the formation of rain in the tropics. Natural precipitation processes are shown schematically in Fig. 3.16.

3.6.3 Artificial modification of rainfall

The Wegener-Bergeron process and the coalescence process depend on the presence of relatively rare ice nuclei and large water drops respectively. If the appropriate nuclei could be artificially introduced into those clouds which had a deficiency of them then it might be possible to trigger the rain forming processes. Mason (1971) gives a good account of the history of man's attempts to stimulate rainfall, and detailed references will be found in his book. In brief, Findeisen in 1938 suggested that rain could be produced if artificial ice nuclei were introduced into supercooled clouds. Before this in 1930, the Dutchman Veraart almost certainly made some clouds rain by dropping dry ice (frozen carbon dioxide), and other things, into them. However, his results were received with scepticism by his fellow countrymen because he made extravagant claims and offered no theoretical support for this methods. In 1946 Vincent Schaefer, in

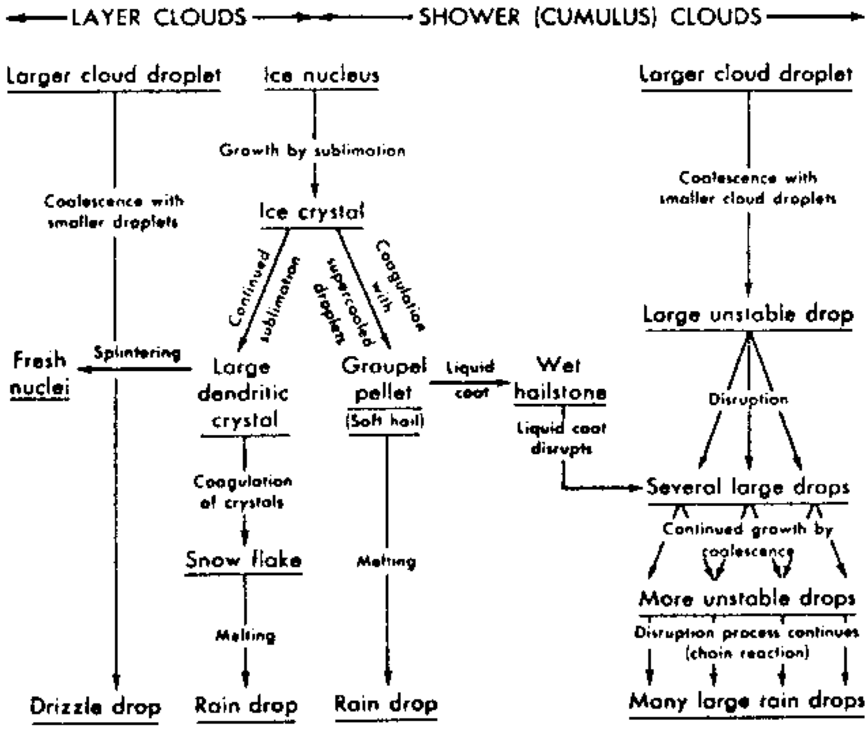


Fig. 3.16. A schematic diagram of natural precipitation mechanisms (From Mason 1957).

America, accidentally discovered that particles of dry ice could produce millions of ice crystals in a cold chamber filled with supercooled cloud. Subsequently, in 1947, he and Irving Langmuir convincingly modified natural supercooled clouds by spreading 1.4 kg of dry ice on them from an aircraft. Not long afterwards Bernard Vonnegut, of the same team, demonstrated that a smoke of silver iodide crystals would accomplish the same result. Silver iodide crystals act directly as ice nuclei because they have a crystal structure similar to that of ice; dry ice on the other hand, being colder than $-70^{\circ}C$, cools some of the air through which it falls to below $-40^{\circ}C$ so forming millions of tiny ice crystals in its wake, each such crystal forming the perfect ice nucleus. When a supercooled cloud is turned into an ice cloud in this way, latent heat of fusion is released and the warmed cloud sometimes grows higher because of its increased buoyancy. This was spectacularly demonstrated by Kraus and Squires in Australia in 1947 when they seeded large cumulus clouds. Six out of eight seeded clouds produced rain-drops but one of these clouds grew rapidly from 7 km to 12 km in the 13 minutes after seeding.

In the mid 1950s several experiments were made to "seed" warm clouds with giant nuclei of one form or another. Fournier d'Albe and collaborators (1955) dispersed salt particles into the air in Pakistan on the assumption that these would then be carried into the clouds by updrafts. Sansom and colleagues (1955) shot salt particles into clouds in rockets, and large water drops (50 to 150 microns radii) have been sprayed into clouds in U.S.A., Australia and Hong Kong (Ramage and Bell 1955). Some of these trials indicated that some warm convection clouds were probably induced to produce rain.

Some of these cloud physics experiments have convincingly demonstrated that both warm and supercooled clouds can be stimulated to rain under certain restrictive conditions. However, this is not the same thing as increasing rainfall at the ground. The difficulty is that, given a certain cloud, it is not yet possible to determine how much rain it would yield naturally, therefore, if the cloud is treated artificially in some way, it is not possible to say whether it produced more or less rain than it would have done if left to itself. In some regions of the world "seeding" experiments have continued over many years in the hope that a sophisticated statistical treatment of the results would produce a decision, one way or another, which would be generally acceptable to both statisticians and meteorologists. A man-made variation in rainfall of about 10% is being sought in a natural variation which may in some cases, amount to several hundred percent. It would seem that some small increase in rainfall may yet be proven in certain parts of the world where freezing nuclei are particularly sparse and supercooled clouds occur frequently. It is important to many water-short countries to know whether rainfall can be significantly increased and under what conditions. Realising that a considerable number of able scientists will be required if this problem is ever to be convincingly resolved the W.M.O. decided in 1975 to co-ordinate an international attack on the problem by launching experiments in selected regions where climatic conditions should be favourable for such experiments.

Sophisticated experiments on tropical cumulus clouds are already being undertaken, as part of Project Stormfury (sect. 18.3) in an attempt to better understand the basic principles of cloud growth and precipitation processes so that they can be simulated in computer models. Mankind will reap more benefit from these scientific experiments than from the hundreds of publicised and sensational "rainmaking" operations that have been carried out in the last two decades.

3.7 Winds

The atmosphere never stops moving and, indeed, this is fortunate for were it otherwise we would all soon be suffocated by our own exhalations. The air occasionally appears calm but motion can readily be detected if sufficiently sensitive methods are used. The atmosphere is probably most nearly still in shallow valleys during the winter in anticyclonic conditions. If such near calm conditions persist for several days they are frequently accompanied by fog and, in industrial areas, the concentration of airborne pollutants may then become so severe as to cause increased mortality in both man and animals as occurred, for example, in Donora, U.S.A. in 1948 and in London in 1952. Winds are therefore necessary to disperse pollutants but, they are also necessary to maintain moderate air temperatures over the world. What causes the wind?

Anyone who has lost a balloon whilst it was being inflated will realise that air moves rapidly from the region of high pressure inside the balloon to the lower pressure outside. Forces are necessary to cause air to move and the force that causes winds is known as the "pressure gradient force". In Fig. 3.17 it is seen that the pressure gradient force is directed perpendicular to the isobars towards low pressure; this direction is also the one in which the pressure changes most rapidly with distance. The rate of change of pressure with distance is known as the "pressure gradient" and it determines the magnitude of the pressure gradient force. When isobars are close together the gradient - by analogy with contours on a map - is said to be large. From the moment that an air particle begins to accelerate down a pressure gradient several other forces begin to assume importance.

Because of the rotation of the earth, bodies moving freely in a horizontal direction appear - to an observer on earth - to be deflected from a straight course everywhere except at the equator. To understand why this is so, consider an observer standing at the north or south pole, he would rotate once in 24 hours about his own axis. At the equator there is no such rotation in the horizontal plane, an observer there rotates only in a vertical plane perpendicular to the earth's axis, and so goes head-over-heels once in twenty four hours. The rotation of a standing observer about his vertical axis decreases smoothly with latitude from the polar maximum, to one half the value at latitude 30^{o}N, and zero at the equator. The head-over-heels rotation decreases similarly, from its equatorial maximum to zero at the poles, but the latter rotation has no significant effect on winds and will not be considered further.

People, not at the equator, often find it difficult to appreciate that the ground on which they stand rotates about the local vertical; there is no object, stationary relative to the centre of the earth, against which they can measure this local rotation. The sun, however, approximates such a reference and the shadow it casts on a sundial, outside the tropics, shows that the dial does indeed rotate about the local vertical.

The many air particles comprising the wind move freely over the earth and, therefore, do not participate in its rotation. However, to an observer rotating about his vertical axis with the surface on which he stands a steady wind maintaining a straight path in space appears to be deflected from its path by some force. In the northern hemisphere the observer (and sundial) rotates in an anticlockwise direction (seen from above), to him therefore, winds will appear to be deflected in the opposite direction that is, to the right. This apparent acceleration to the right due to the rotation of the earth is called the "geostrophic acceleration" or "Coriolis acceleration" and is regarded as being caused by an apparent "force" which affects all freely moving bodies when their motion is referred to the earth. In the southern hemisphere the rotation is in the opposite direction and the deflection is therefore in the opposite direction (to the left). To avoid frequent qualification we will, from here on, be concerned only with conditions as they are found in the northern hemisphere: the reader may work out the corresponding conditions in the southern hemisphere if he so wishes.

A thorough and more general demonstration of the Coriolis acceleration requires the use of mathematics but the unconvinced reader can consider the following practical demonstration of the effect. A missile once fired has no forces acting on it to make it participate in the earth's rotation, however, during the projectiles free flight the earth rotates so that, on the missile arriving at the point of aim, the target on earth will have moved. The gunner, sees the error as being caused by some force deflecting his projectile. In long-range artillery and rocketry therefore, allowance has to be made for the Coriolis force. It is fortunate for golfers that both the brief time of flight of a golf ball and the short distances involved make the deflection too small to be identified amongst the pertubations due to wind, spin and incompetence.

Readers will doubtless have been told, at sometime, that the vortices formed as water empties from a bath rotate anticlockwise north of the equator and clockwise to the south. Some claim further, that such a vortex on board ship changes direction as the equator is crossed! The true facts are that, away from the equator, the Coriolis acceleration would induce an appropriate rotation in water moving towards the outlet were it not for the fact that other small, residual motions are usually dominant. By leaving water to stand in a large circular dishpan for 24 hours or more to become still, by replacing the plug by a long tube and remote tap and by protecting the tub from vibration and drafts then, the water vortex will always twist in the appropriate direction in middle and higher latitudes. Near the equator the Coriolis acceleration is so weak -there is little rotation in the horizontal plane - that its effects defy detection amongst other small but nevertheless stronger residual motions. Claims that, on crossing the equator, bath tub vortices change their direction of rotation because of Coriolis force are, therefore, completely unfounded. It is possible that vortices might, on occasion, change direction at such times but the cause cannot be attributed to the rotation of the earth.

The Coriolis force is a major factor in the large scale motion of the atmosphere. Its magnitude depends on the speed of the wind v and the rate of the earth's rotation in the horizontal plane. The latter is dependent on latitude ϕ and the angular velocity of the earth ω . In mathematical terms the geostrophic force on unit mass is given by:

$$\text{Coriolis force} = 2\,\omega v \sin\phi \quad \text{.........................} \quad (3.1)$$

and it disappears at the equator ($\phi = 0^{o}$) where Sin $\phi = 0$.

From the foregoing discussion it will be appreciated that as particles of air are accelerated down the pressure gradient by the pressure gradient force, the Coriolis force causes an apparent deviation to the right. The wind is accelerated by the pressure gradient force until a velocity is reached such that the Corolis force equals the pressure gradient force, the wind then comes into balance blowing along the isobars with a steady speed known as the "geostrophic wind", Fig. 3.17(a). If the air is slowed down - i.e. smaller v - the deflecting force will decrease and the air will then turn towards the left, that is, towards lower pressure under the influence of the now dominant pressure gradient force. The latter accelerates the air again increasing V and the geostrophic force, until balance is again achieved.

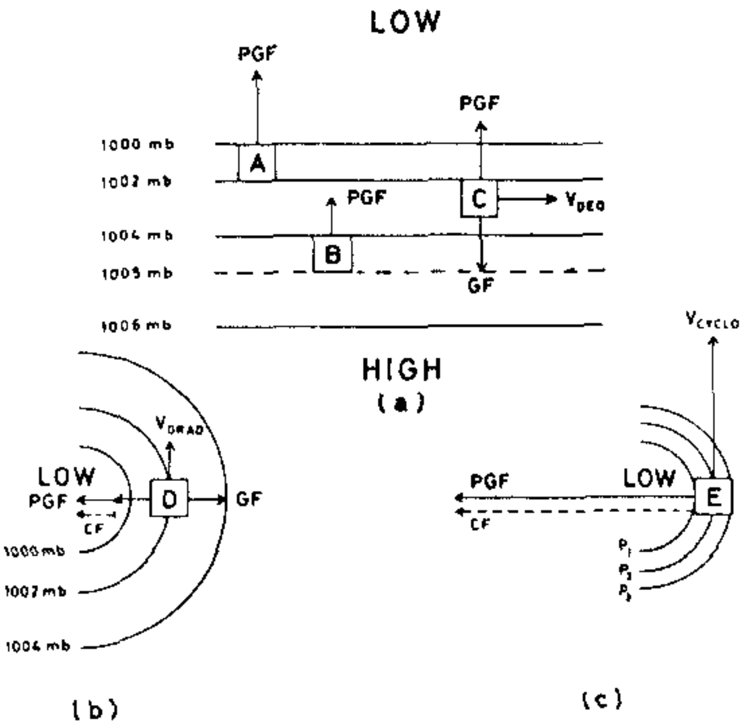


Fig. 3.17. The cube of air A, has a pressure of 1002 mb on one face and 1000 mb on the one opposite, it therefore experiences a net pressure gradient force, PGF, equal to the pressure difference of 2 mb multiplied by the face area. The same cube at B, where the pressure gradient is one half that at A (isobars twice as far apart) experiences only half the PGF. At C, the cube moves with the geostrophic wind speed so that the consequent Coriolis or geostrophic force GF balances the PGF. In (b) an additional centripetal force CF is required to keep the cube on a curved path. At the speed of the gradient wind the PGF exceeds the geostrophic force GF by an amount CF adequate to maintain curved flow. In (c), not to scale, GF is neglected as it is small in comparison to the large PGF (due to the ~~steep~~ large gradient) and large CF (to hold fast moving air on the sharply curved path). The wind calculated on this assumption i.e. that PGF is equal to CF, is known as the "cyclostrophic wind".

In nature, isobars are seldom straight and for balanced flow around a curved path it is necessary to provide a net force towards the depression centre - a "centripetal force" - to constrain the air to follow the curve. Winds under such circumstances are called "gradient winds" and for flow around an area of low pressure, such as a tropical depression, the velocity of the wind adjusts so that the geostrophic force is less than the pressure force by the amount necessary to provide a centripetal force adequate to maintain curved flow, Fig. 3.17(b). The force of friction slows the wind near the earth's surface thereby reducing the deflecting force to less than that needed for balance, the wind than blows across the isobars towards lower pressure. The frictional effect decreases with height and eventually becomes so small as to permit the wind to flow along the isobars as required for balance; the level at which this occurs, "the gradient level", is usually about 1 km above the surface and marks the top of the "friction layer". Cumulus and other low clouds usually move with the wind at the gradient level and as we look downwind they will be seen to move to the right (N. hemisphere) of the flow at lower levels. Over the tropical oceans, on average, the change of wind direction between the surface and 1 km is approximately 10° (Fig.) but on occasions the deviation can vary widely from this average. It is the force of friction which, in typhoons and cyclones generally, causes the low level winds to spiral across the isobars at an angle of 30 to 40 degrees towards the centre, Fig. 1.4 The air in the friction layer therefore accumulates (or converges) towards the centre where it rises to produce clouds and rain. Conversely, the wind blows away from the centres of high pressure (i.e. towards low pressure) so causing the overlaying air to sink -subside - to replace that which is lost; the sinking air warms and dries and, in general, produces clear, fine weather.

Near the centres of typhoons the pressure gradient is very large indeed and the air there moves rapidly along a highly curved path; under these circumstances both the pressure gradient force and the centripetal force are so large as greatly to exceed the geostrophic force. The contribution of the latter, being relatively small, can be neglected when calculating the wind speed; such a wind in which the centripetal force necessary to maintain curved flow is provided entirely by the pressure gradient force, Fig. 3.17(c), is known as the "cyclostrophic wind" and it is independent of the geostrophic force. This approximation is only true for the highly curved, high speed winds near the centre, further away where the wind speed is lower and the path less curved the geostrophic force is important. It follows that, if a typhoon or tropical storm were small enough, its winds could circulate in either a clockwise or anticlockwise direction maintaining cyclostrophic balance as in waterspouts and dust devils (sect. 3.13); it would then be able to cross the equator and move some distance into the other hemisphere. In practice, however, tropical cyclones are too large for the geostrophic force to be ignored and they have never been observed to cross the equator although it is incorrectly stated otherwise in some texts. Indeed, we can go further and state that the geostrophic acceleration arising from the earth's rotation is necessary for both the formation and maintenance of tropical cyclones.

The geostrophic, gradient and cyclostrophic winds refer to steady state flow and are not valid in changing conditions at which time the velocity of wind can depart considerably from that indicated for the steady state. For example, when the surface pressure changes with time it is possible to draw lines - "isallobars" - which join those places on a weather chart at which the pressure is falling or rising at the same rate. Under these conditions it is found that an additional component of wind - the "isallobaric wind" - blows across the isallobars from areas of rising pressure towards areas where the pressure is falling or not rising so quickly. Furthermore, the weak Coriolis force found in the tropics is - in certain situations - readily overwhelmed by other forces so permitting the surface wind to blow across the isobars at right angles - down the pressure gradient - and with a speed other than that of the gradient wind. This fact is not fully recognized in many text books which contain statements which cannot be reconciled with the facts depicted in Figures 3.18 and 3.19. The flow of air directly across the isobars, towards low pressure, as shown in these figures occurs when air arrives at a place with a velocity less than that required to balance the prevailing pressure gradient - this situation can arise in several ways. In Fig. 3.18 cold air moving southward subsides and, because it originated at higher levels in a westerly current, arrives near the surface with a velocity too small for balanced flow in the large pressure gradient in its new environment. Its relatively low velocity results in an inadequate geostrophic force for balanced flow and the air is therefore driven towards

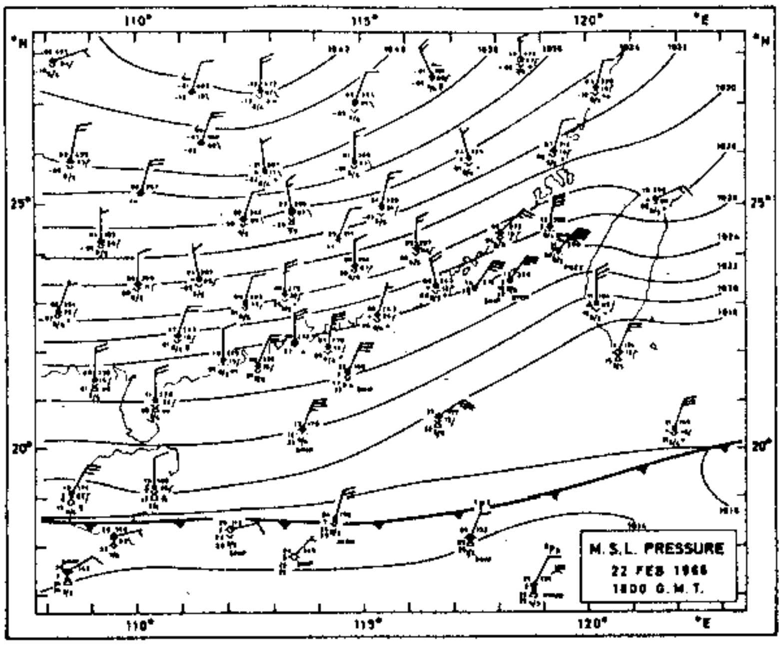


Fig.3.18 During fresh surges of the N.E. monsoon surface winds over land often blow across the surface isobars at right angles. On this occasion, at Hong Kong, at a height of 1.1 km the wind was blowing along the isobars (080°) at 6.5 m/s and surface pressure was rising at a rate of 1.1 mb/h.

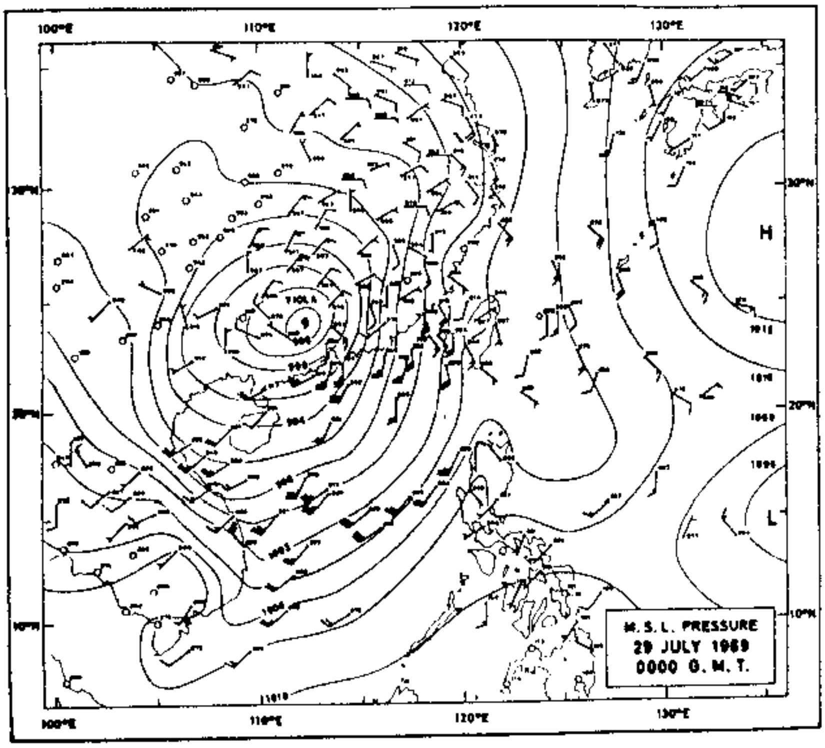


Fig. 3.19 Surface winds over land blowing directly towards the centre of typhoon Viola as it weakens after crossing the coast.

low pressure by the pressure gradient force and isallobaric force against the restraining force of friction. In Fig. 3.19 as typhoon Viola crosses the coast increased friction over land slows the surface wind so that both the centrifugal and Coriolis forces - which depend on wind velocity - weaken and become unable to balance the pressure gradient force which then drives the air towards low pressure against friction. The true relationship between wind and pressure in the tropics is not known (Frost & Stephenson 1965).

In and near the tropics the cyclostrophic wind is a fair approximation to the maximum winds in a typhoon and, in this application, is usually within about 5% of the gradient wind. In temperate latitudes the geostrophic wind (the straight flow case) is used, in some applications, as an approximation to the gradient wind; this is not acceptable in the case of typhoons because of the high speed and curvature of the flow. For example, a moderate typhoon at 20°N might contain a maximum pressure gradient of 1 mb/km; the corresponding geostrophic wind is 2 000 m/s! If the radius of the maximum winds is 20 km then the cyclostrophic wind will be 44.7 m/s and the gradient wind 44.6 m/s. In this instance the centrifugal acceleration is 45 times as great as the Coriolis acceleration. Nomograms for determining the geostrophic, cyclostrophic and gradient winds in typhoons are shown in Fig.5

In summary we can say that the greater the pressure gradient the higher the wind speed; the strongest winds will therefore be found where the isobars are closest together. In the tropics over the sea, on average, the gradient wind level is near 1 km. The surface wind has a speed about 25% less than that at the gradient level and the frictional turning is about 10° (Atkinson & Sadler 1970). However, the speed reduction and the turning are much greater over rough terrain - such as a large, tree covered island - and can amount to 50% and 50° respectively. Large departures of the wind from that to be expected from balanced flow often occur in the tropics.

Check with chap. 5

3.7.1 Measurement of wind speeds

The surface wind is defined as the 10-minute mean wind at 10 m above the sea or over open flat ground. Its speed is measured by anemometers, some of which use a propellor or a 3-cup windmill to turn an electric generator. These instruments purport to record wind speed variations which last a second or more and the anemograms which they produce show that the wind is seldom steady (Fig.). Short period increases in wind speed are called "gusts";sudden, sustained increases lasting for some minutes, known as "squalls", are usually associated with passing convective showers or thunderstorms. Sometimes the showers or thunderstorms are arranged along a line which is then known as a "squall line".

Hydrogen or helium filled balloons are released at many stations at midnight and midday GMT; they are tracked by theodolite or radar to determine wind velocities in the upper atmosphere. At most stations four balloons are released each day, the extra ascents being made at 0600 GMT and 1800 GMT. The main ascents carry radiosondes which report back by radio the pressure, temperature and humidity of the environment to enable the heights of pressure surfaces to be calculated.

3.7.2 Representation of winds

Surface and upper winds are plotted on appropriate charts as arrows flying with the flow, the number of barbs indicating the speed. The charts are then completed by drawing lines which indicate the direction of the instantaneous flow; all wind arrow should therefore be tangents to these lines which are called "streamlines". Because there is no quantitative relation between streamlines and wind speed, additional lines called "isotachs" are usually drawn; these connect places having the same wind speed (Fig. 3.20a). Streamlines give the direction of the wind at the time for which the chart is valid, it is important to realise that they do not represent the historical path, or "trajectory", of the air unless the streamlines are not changing with time.

Winds can also be represented by isobars on surface charts and contours (sect.3.3) on upper-air charts. From the gradient level upwards the wind for all practical purposes is taken as blowing along the isobars (or contours) and its speed is given by their spacing and latitude, according to the gradient wind relationship. The surface wind is derived from the gradient wind by slowing it down and turning it towards low pressure, as previously discussed.

In the tropics there are only small differences in the height of the pressure surfaces over large distances (Fig. 3.21d). These small spatial height differences are difficult to measure because they are comparable in size to the errors of measurement and the local variations such as might be found between ascents made in a cloud and those in the surrounding clear air. Instrumental errors in the heights of pressure surfaces in the tropics are often so large that some have to be ignored to permit the drawing of reasonable contours on both synoptic and averaged charts (Frost 1970). Contours are therefore of little value for forecasting in the tropics (Ramage 1964, Riehl 1966). Mean heights are better related to mean winds than are their synoptic counterparts nevertheless, areas can be found where there is no gradient of height but a strong wind (e.g. central China, Fig. 3.21d) and elsewhere long period means of the height of the 200 mb surface at neighbouring stations using the same instruments are inconsistent with the wind flow and each other. The same difficulties are encountered in other tropical regions. Frost (1970) examined the relationship between the wind and the contours at 200 mb over tropical Africa and he writes:

" It is most unfortunate that even for the most intense wind circulations in the tropics and subtropics i.e. the westerly jet, the easterly jet and the duct, the height differences between the equator and 15°N are of the order of $\pm$ 30 to 40 gdm whilst with the radiosondes at present in use in Africa the monthly height differences have systematic errors of $\pm$ 50 gdm whilst the daily height differences appear to have systematic and casual errors of $\pm$ 70 gdm. Even however if all countries in Africa adopted the same type of radiosonde, and one which is better than any at present in use, and used common procedures it still seems that very little practical use could be made of radiosonde observations for contour analysis for, apart from the small height differences which would be observed, as first pointed out by Crossley (1948) and, as

(1) demonstrated by the calculations .... the height differences by themselves are clearly insufficient to determine the motion of the air."

For these and other reasons streamlines are usually used to depict the wind flow between latitudes 30°N and 30°S, for most purposes, in preference to contours or isobars.

3.7.3 Distribution of winds

Except in the tropics, the mean surface air flow should show some broad relationship to the mean surface pressure distribution; comparison of Figs. 3.3 and 3.22 show that this is indeed so. In winter air flows out of the Siberian anticyclone as the northeast monsoon. In summer, the flow over S.E. Asia is reversed as the winds blow around the monsoon lows, over north India and south China, as the southwest monsoon. Over the North Pacific Ocean, the Pacific anticyclone and its ridge to the westward are evident in both months (Fig. 3.22) but are considerably more developed in August. The equatorial trough of low pressure, marked as a dashed line in Fig. 3.3, is seen in Fig. 3.22 to be associated with areas in which airstreams flow towards one another. Note that this line - a line of confluence - migrates during the year with the equatorial trough although this movement is mainly confined to the eastern hemisphere. The winds used to draw the streamlines in Fig. 3.22 are vector mean winds and are a measure of the total flow at the point in question. They are taken from the atlas by Crutcher and Davis (1969) which is based on 20 million ship weather reports from the last 100 years.

Selected charts of the mean wind flow at upper levels during February and August are shown in Figs. 3.23 and 3.21. The Pacific anticyclone and ridge are evident at all levels and the latitude of the

(2)

ridge changes with height and the time of year. It will be seen from Fig.3.21 and 3.23 that at 140^{o}E, in February, the ridge slopes southward with height whereas in August it slopes towards the north. This remarkable change, is shown in greater detail in Fig. 3.24 where the mean latitudes of the ridge at each level for each month are depicted. As can be seen in Fig. 3.25 the ridge is furthest north, at most levels, in August. In general, it is furthest south in February, as are the circumpolar westerlies, but at levels above 500 mb the ridge does not reach its most southern latitude until May. The ridge at the highest level shown in Fig. 3.25 (200 mb) moves through a greater range of latitude during the year than at other levels; at the surface the ridge moves the least distance. The early or late migration of the ridge at 200 mb has a dramatic influence on the weather over Far East Asia (Bell 1976) and on the development and movement of tropical cyclones. This arises because the 200 mb ridge is associated with the direct cell of the general circulation (the Hadley cell - a meridianal circulation) and its position at any time is related to the broadscale flow patterns over the globe. The Pacific high at 200 mb is seen to be centred over the central tropical North Pacific Ocean in February and to more towards the west-northwest reaching its highest northerly latitude over the Himalayas, in August. The strong westerly winds at this level (200 mb) are seen to extend as far south as North Luzon in April but they retreat northwards during May so that easterly winds cover the tropics by June. These easterlies attain speeds of more than 25 m/s - in the mean - over Sri Lanka during August. The jetstream in the westerlies which is located south of the Himalayas in April moves to the north east of the massif by June; this move is associated with the onset of the southwest monsoon between the surface and 500 mb.

Below 200 mb the flow pattern is more complicated. At 500 mb the circumpolar westerlies, found as far south as 15^{o}N in winter, retreat northward and give way to the southwest monsoon by June which, in turn, is replaced by easterlies on the south side of the Pacific ridge by August. The circumpolar westerlies return to low latitudes by October.

At the gradient wind level (1 km) the northeast monsoon can be clearly seen over China and the Western Pacific in the charts for October, December and February, and the southwest monsoon flow is best developed in the charts for June and August.

There are many features to be seen in Fig.3.25 but we will mention only one more. In February the equatorial trough at 1 km is too far south to be seen on the chart but it can be seen at the bottom of the chart for April. Note that the winds over the trough at 200 mb are easterly; this is a feature common to the charts for all months.

Higher in the troposphere the dominating feature is the broad belt of circumpolar westerly winds, it is best developed in the winter hemisphere when it broadens, increases in speed and extends equatorward into the tropics. The part of this circulation which affects the Western Pacific is shown in Fig.3.21. For many years it has been known that the energy to drive the general circulation comes from the sun. Over a year the tropical belt receives more radiant heat than it looses and, conversely, the Polar regions radiate more heat into space than they receive from the sun. The temperature in the tropics would therefore rise to a new level of equilibrium, and that at the poles would fall, were it not for the atmospheric circulations which carry heat from the tropics to the poles and so maintain the balance. Our knowledge of the details of this circulation, their relative importance and the factors which control them is still far from complete. It is relatively easy to conceive of mechanisms for carrying heat from the tropic to the poles but it is not obvious why it takes place at the observed rate. An increased rate of heat transfer would cool the tropics and warm the poles, conversely a slowing down of the process would produce a new ice age in temperate latitudes and warmer conditions in the tropics. The circulations of the ocean carry approximately 10% of the total poleward heat transport. On average the troposphere itself looses heat by radiation at a rate which would cool it by $1^{o}C$ to $2^{o}C$ each day if it did not receive heat from the earth.

Palmen and Newton's (1969) model of the meridional and vertical components of the general circulation is shown in Fig. 3.20. It was originally thought that the general circulation was a simple heat engine with warm, rising air at the equator and cooler, subsiding air at the poles; although it is now known that the circulation is more complicated, there is nevertheless a direct thermal circulation, or Hadley cell, in our region of interest that is, between the equator and the subtropical high pressure belt. The earth's surface (the heat source) warms and moistens trade wind air as it moves equatorward to rise in many large cumulonimbus clouds in or near the equatorial trough (the doldrums Fig. 3.20). The air, now at relatively high pressure moves northward loosing heat by radiation (heat sink) and so becomes cooler and denser and subsides in the subtropical high pressure belt to maintain the anticyclones there. Some of the air then returns to the equatorial trough as the trade winds.

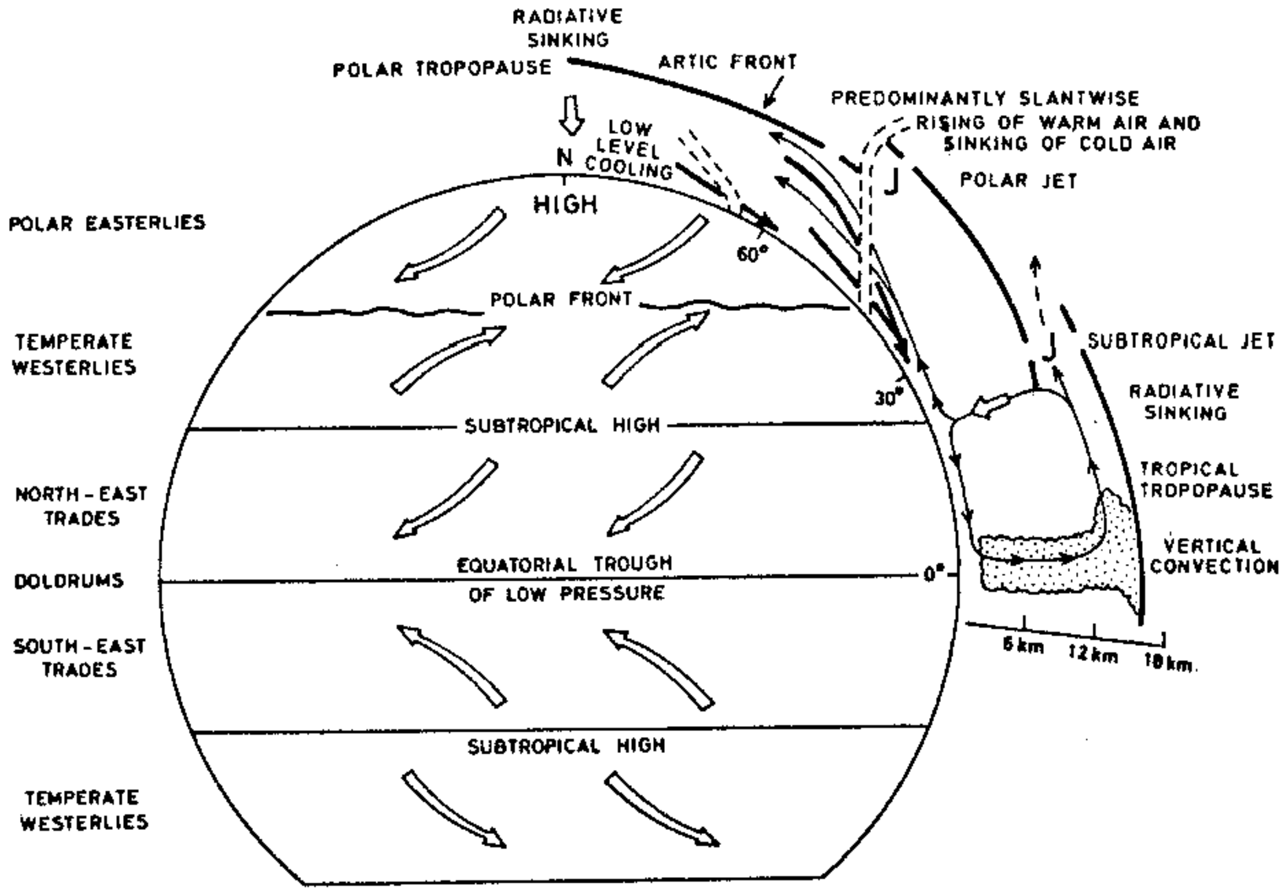


Fig. 3.20. A schematic diagram of the main areas of high and low pressure and the major wind systems of the world which form part of the general circulation. The meridional flow in the vertical plane in winter is after that proposed by Palmen and Newton (1969).

3.8 The general circulation

The charts of mean surface pressure, mean contour heights or resultant winds for individual months or seasons, provide information on the broad scale air motions over the globe without the complications caused by the short-lived weather systems which are seen on daily weather charts. Figure 3.10 shows schematically the broad features of the mean global surface pressure distribution and resultant airstreams which comprise the surface component of what is known as the "general circulation". The following features of the pressure distribution are found in both hemispheres:-

1. The near equatorial trough of low pressure between 10^{o}N and 10^{o}S.
2. The subtropical high pressure belt between latitudes 10^{o} and 40^{o}S.
3. The temperate latitude low pressure belt between latitudes 40^{o} and 70^{o}S.
4. The anticyclone centred poleward of 70^{o}.

These belts of high and low pressure are not continuous but consist of a number of separate semi-permanent anticyclones or areas of low pressure some of which change position during the year in response to the seasonal march of the sun and the differential heating of land and sea. The subtropical high-pressure belt contains the Pacific anticyclone and its ridge which extends westward towards China. These important semi-permanent features can readily be identified on daily weather charts and they are related to the formation and movement of typhoons as also are the trade winds and the near equatorial trough.

At the equator the earth has a speed of 464 m/s towards the east; and air rising there will share this speed. As the upper return flow moves northward it will arrive at places closer to the earth's axis of rotation and so will speed up - like a ballerina speeding a pirotte by lowering her arms to bring them closer to her axis of rotation - and produce strong westerly winds. This is a manifestation of the principle of the conservation of angular momentum.

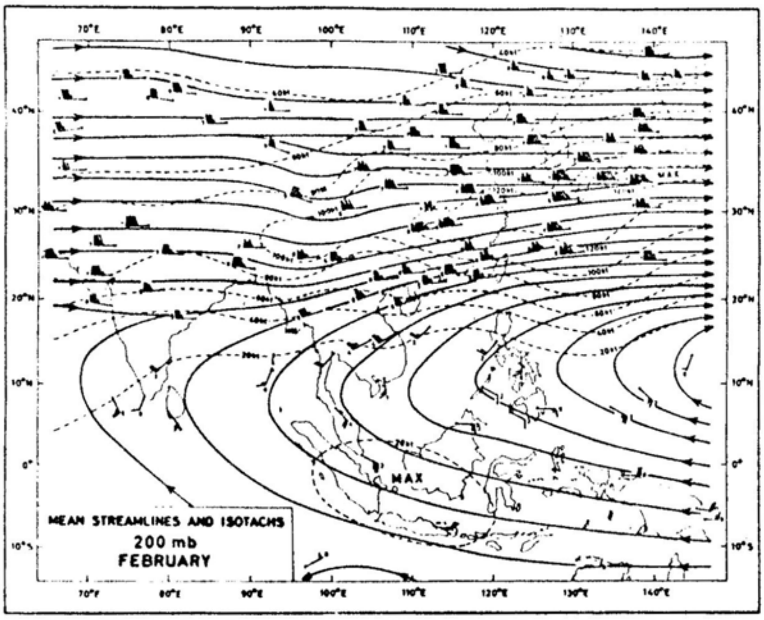
In these westerly winds, near the two places marked J in Fig. 3.20, X the wind speed is markedly higher than that a few tens of kilometres to the north and south. These two streams of strong winds meander around the world and are known as "jet streams" and they occur in both hemispheres.

3.8.1 Jet streams

In the mid-1930s the famous Swedish meteorologist, Carl-Gustav Rossby studied the Gulf Stream, a narrow ocean current with a speed of a few metres per second embedded in water moving at approximately one hundredth of this speed. He sought similar streams in the atmosphere but was unable to find them because the density of upper-air observations before World War II was inadequate for this purpose. The conditions did not improve materially until, in the early 1940s, more information became available and Rossby and his associates were able to find narrow high speed airstreams in the westerlies at a height of about 12 km. In 1947, they described these fast-flowing currents, thousands of kilometres in length, a few hundred kilometres in width and a few kilometres in depth and named them "jet streams".

Since 1911, balloon flights had indicated that exceptionally strong westerly winds could sometimes be found in the upper-air and, further, movement of high clouds had shown that the wind in strong airstreams often changed speed markedly in a relatively short distance across the stream. On 19th October 1917 a fleet of zepplins, returning to Germany after attacking England, was widely dispersed over France by a strong northerly current of about 70 knots at a height of 6,000 m and, to even the account, a British bomber force on its way to Berlin on the night of 24th March 1944 was blown to the south of its course by a jet stream; one crew found themselves over Leipzig 160 km south of Berlin. It was not until the latter half of World War II that aircraft capable of reaching the core of jet streams were used in any numbers. In 1944, Boeing B29 aircraft of the U.S.A.A.F. flew from the Marianas to bomb Japan and the crews on one flight found themselves in the embarassing position of being stationary over their targets whilst flying at speeds in excess of 100 m/s. They were, of course, heading into the world's strongest and most persistent jet stream which, in some months, averages over 70 m/s (Fig. 3.21 a). Japanese meteorologists knew about this jet stream early in World War II and they used it to transport balloons carrying fire-bombs to the U.S.A. About 1000 of these balloons successfully crossed the Pacific to America but total secrecy on their landings led the Japanese to believe that the balloons were not finding their targets and they discontinued releasing them. Jet streams have a profound influence on aircraft navigation. In the early 1950s Stratocruisers and other propellor-type aircraft used the Pacific jet stream to help them fly from Tokyo to Hawaii without the need to refuel at Midway Island. These aircraft flew at about 7.5 km which is too low to reach the core of the jet but, nevertheless, they occasionally experienced average tail winds of up to 60 m/s over the whole 5,440 km from Tokyo to Honolulu. In 1952 the first jet airliner, the Comet I, flew in this jet stream from Osaka to Tokyo covering the ground at supersonic speed. When aircraft have to fly in opposition to a jet stream they are routed, whenever possible, to fly below or to one side of the jet stream core. The high speeds of modern jet aircraft make them less sensitive to wind speeds, nevertheless, flights over the 3000 km from Hong Kong to Tokyo, in February, for example, typically take 3 h 10 min northward and 4 h 20 min or about 37% longer to return.

Both the intensity and the relatively fixed position of the jet stream in the Western Pacific is attributable to the effect of the Himalayan Massif. The jet forms over China where warm air from the south of the massif meets colder air from its northern side (Fig. 3.4). Winds over or just to the south of Japan reach over 100 m/s at sometime in most winters. Such strong winds cannot be reliably measured using routine techniques. Balloons are rapidly blown towards the horizon and the resulting small angles of elevation cannot be measured with sufficient accuracy. To overcome this difficulty a special technique was introduced whereby balloons released from Honjo (60 km N.W. of Tokyo) in the winter months after 1951 were observed both from Honjo and from Tateno, a station about 88 km downwind. When balloons were blown too far from Honjo to be accurately tracked they were followed from Tateno. Arakawa (1956) reported that winds at 11 km measured by this relay technique during the first 15 days of February 1953 varied in speed from 70 m/s to 143 m/s and on the 2nd March 1954 a wind of 150 m/s was observed at 12 km before the balloon came too close to the horizon to be reliably tracked. Although winds in excess of 170 m/s are occasionally reported from stations in Japan the inaccuracies inherent in the standard observational method make



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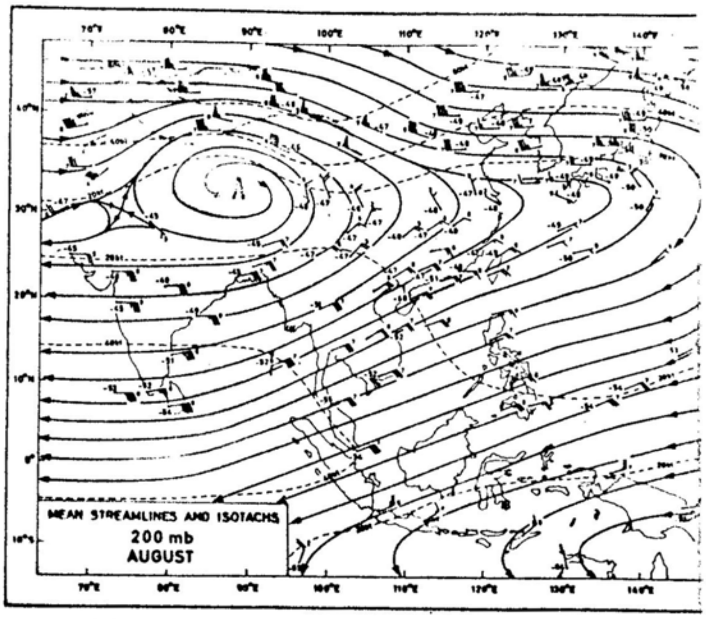

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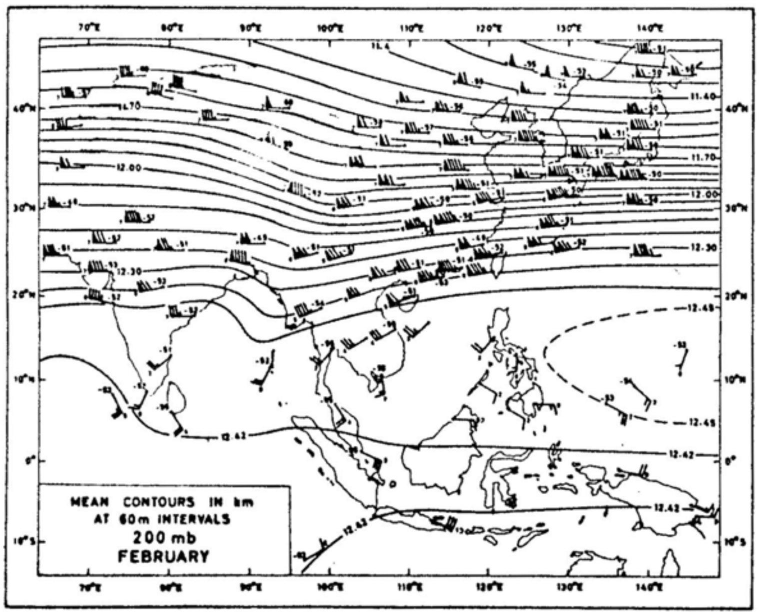

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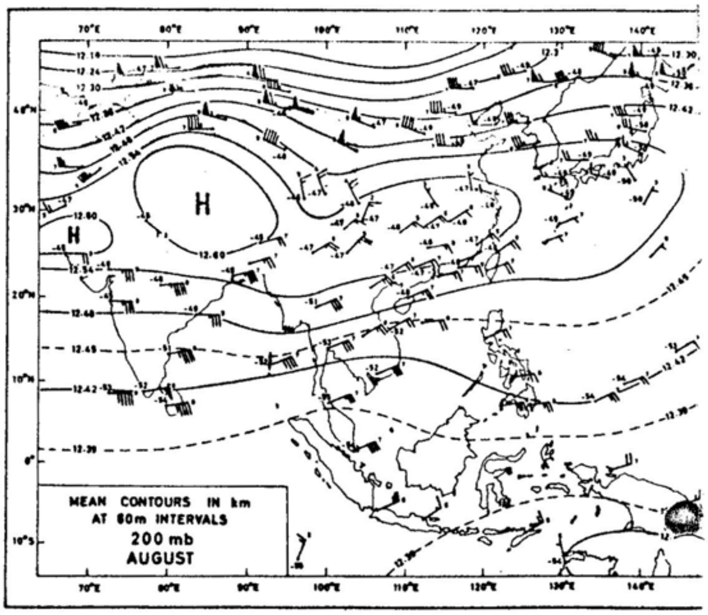


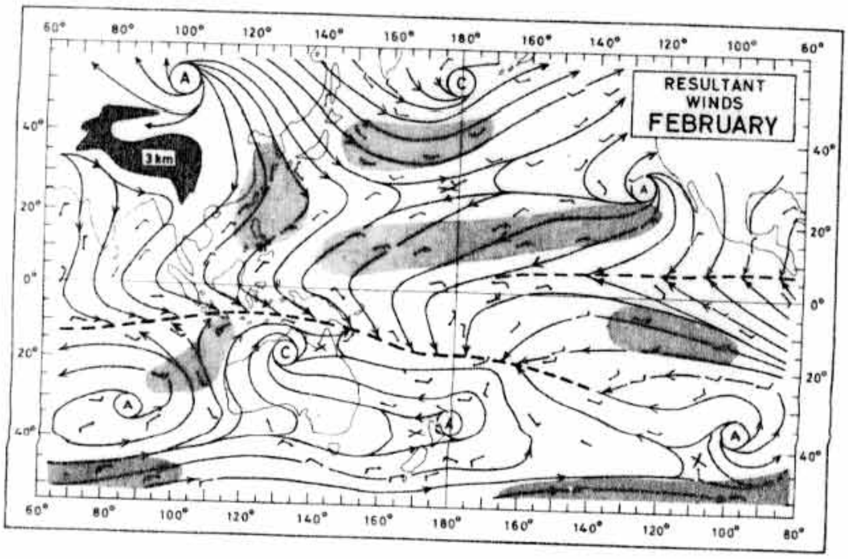
d

Fig. 3.21. Mean streamlines, isotachs and contours over the Far East and Western Pacific at 200 mb during February and August. The arrows fly with the wind, full barbs represent 5 m/s (10 knots) and each triangle represents 25 m/s (50 knots). The direction of the wind to the nearest ten degrees is indicated by the small figure near the barbs, for example, 5 means 50°, 150°, 250° or 350° the appropriate value being obvious from the arrow as plotted. Mean temperatures in °C are shown on the contour charts. That part of the Himalayan Plateau above 3 km is shown shaded.

them suspect. It is generally believed that the upper limit of the wind speed in the jet core is probably around 170 m/s (330 kn).

Jet streams frequently cause clear air turbulence which is occasionally so violent as to be a hazard to flight. We do not yet know enough about this phenomenon but we do know that, away from mountains, turbulence is most frequently found on the low pressure (cold) side of the jet stream in association with strong shear of the wind near the jet core. In particular, clear air turbulence appears to be likely if the horizontal shear is greater than 30 m/s in 200 km (15×10^{-5}/s) or the vertical shear is greater than 50 m/s over 6 km (8.3×10^{-3}/s). Arakawa (1959) found a wind shear in the vertical of 25 m/s in 390 m (0.064/s) above Tateno on 6th March 1954. The jet stream is an important factor in the development, movement and intensity of extra-tropical depressions and, sometimes, of tropical cyclones.

Although we have so far spoken only of jet streams in the westerlies they also occur in the tropical easterlies of the summer months reaching their highest speeds in August over the southern tip of India (Fig. 3.21 b and d).



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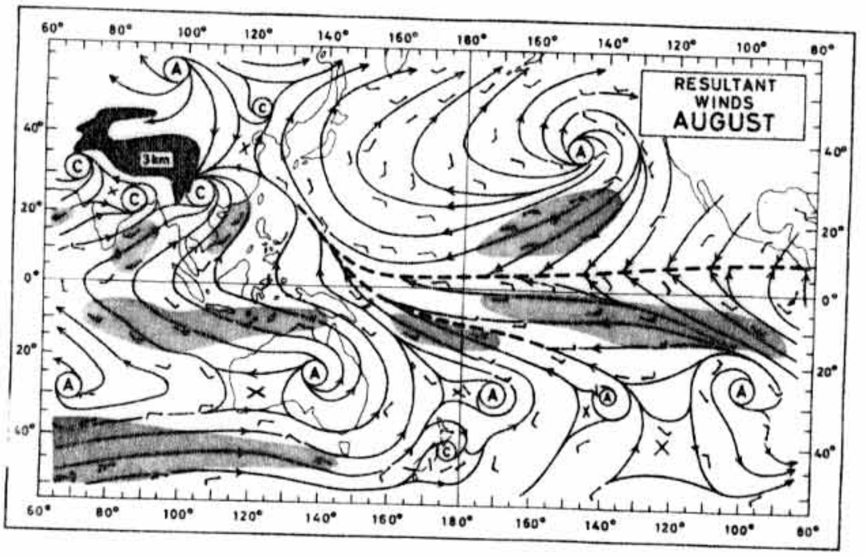

(b)

Fig. 3.22 Resultant (vector mean) surface winds in (a) February and (b) August with streamlines. The centres of anticyclonic flow and cyclonic flow are marked A and C respectively. The dashed lines indicate lines of confluence and the light shaded areas indicate regions where the wind speed exceeds 15 knots. The Himalayan Massif is shown in darker shading. (The winds are from Crutcher and Davis 1969)

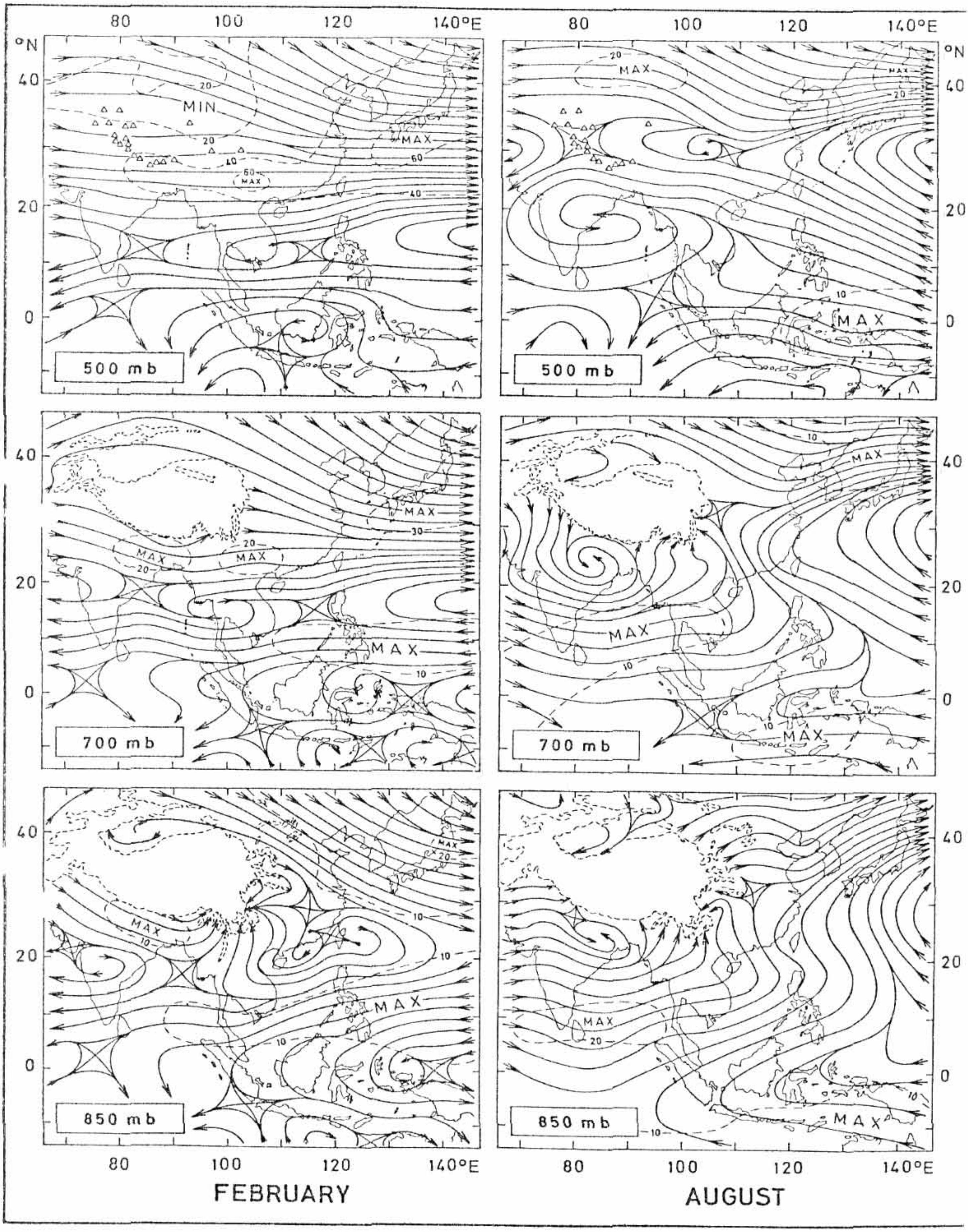

Fig. 3.23. Streamlines for three levels in February and August. Those for 200 mb in the same months are given in Fig.3.21. Isotachs are labelled in knots. (After Chin and Lai 1974).

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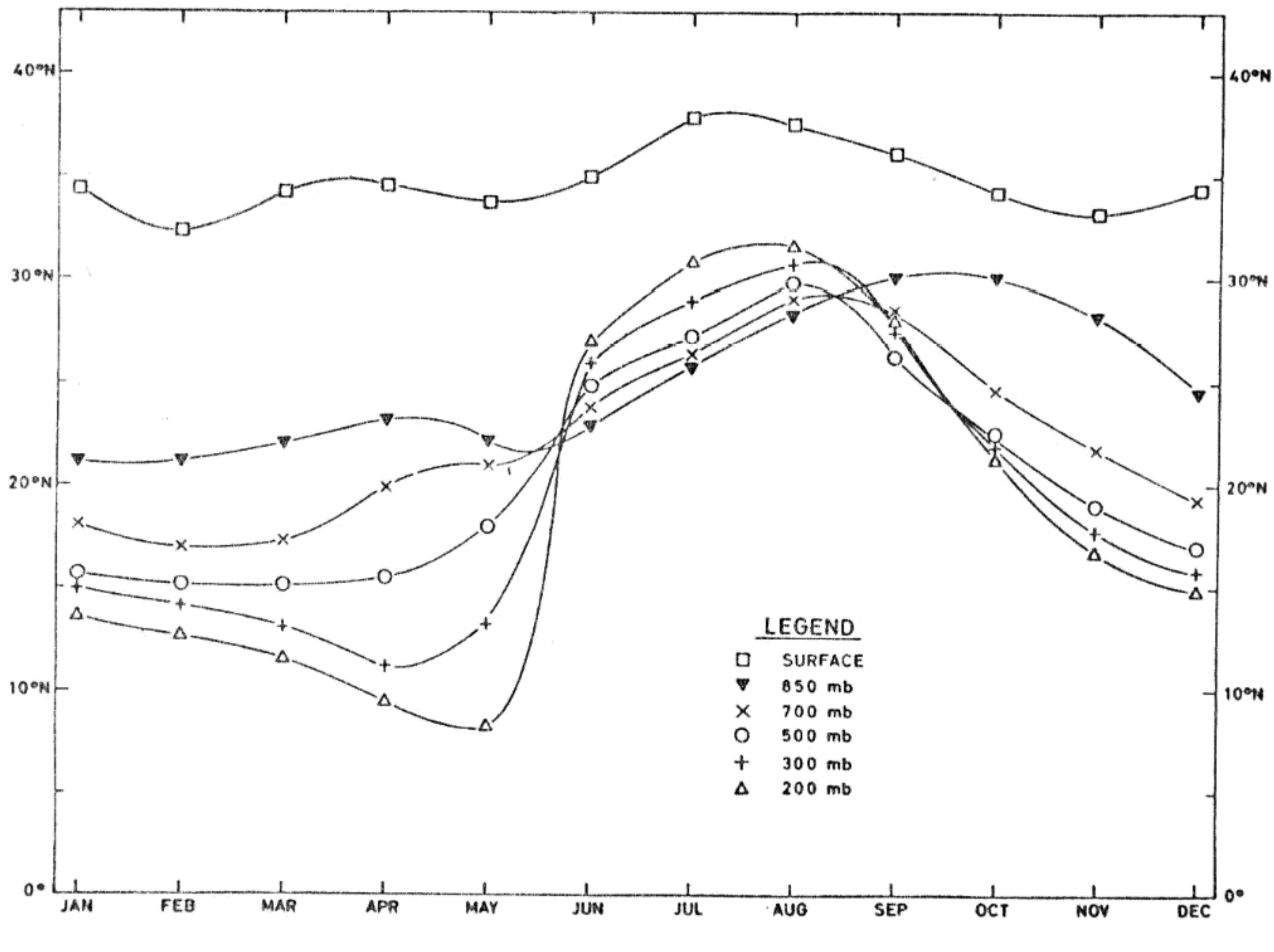

Fig. 3.24. The mean latitude of the Pacific ridge along the meridian 140°E is shown here for each month and six levels.

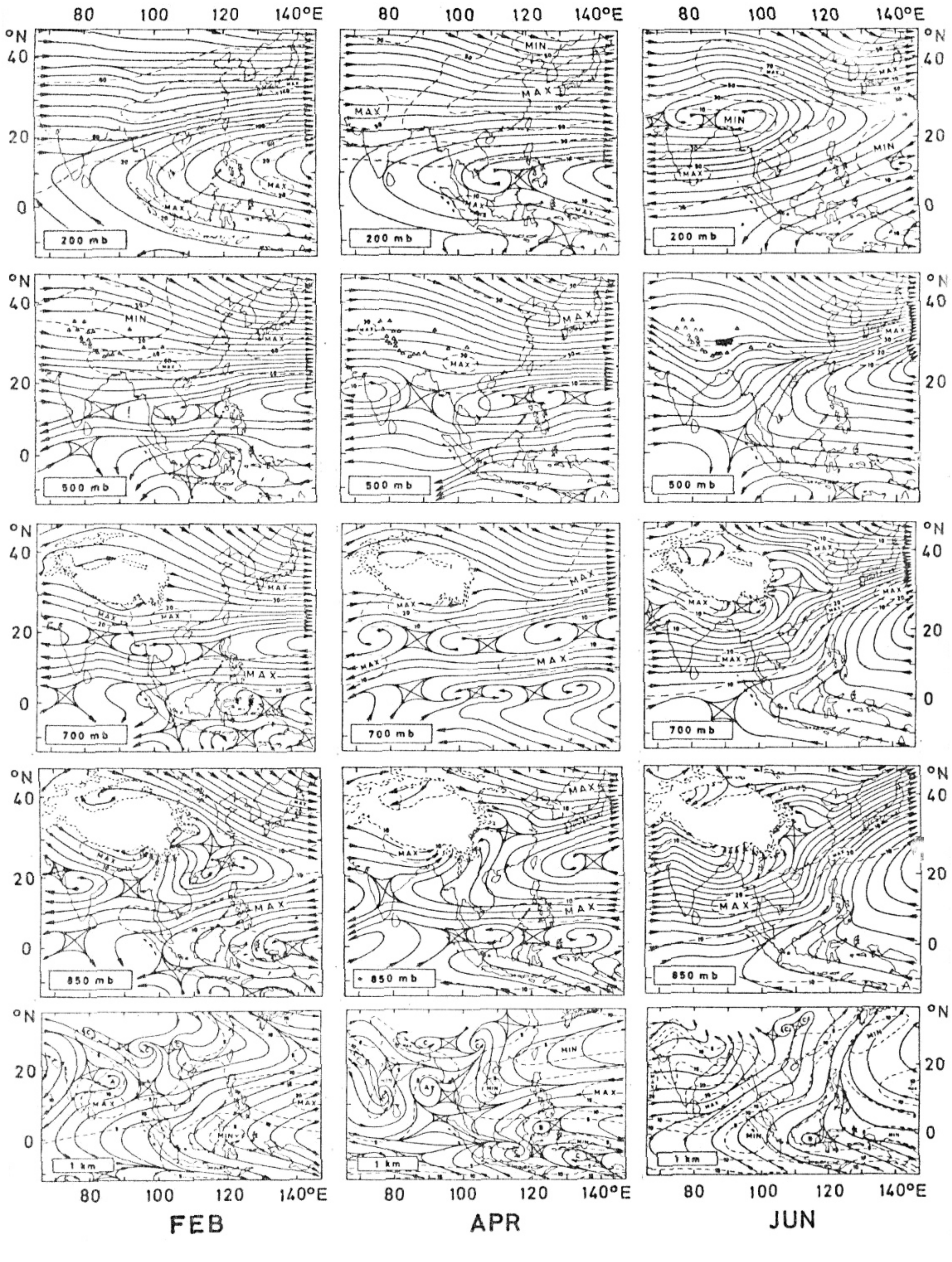

Fig. 3.25. Mean streamlines and isotachs (in knots) at five levels in six different months. (The gradient level winds - 1 km - are after Atkinson and Sadler 1970, the others are after Chin and Lai 1974).

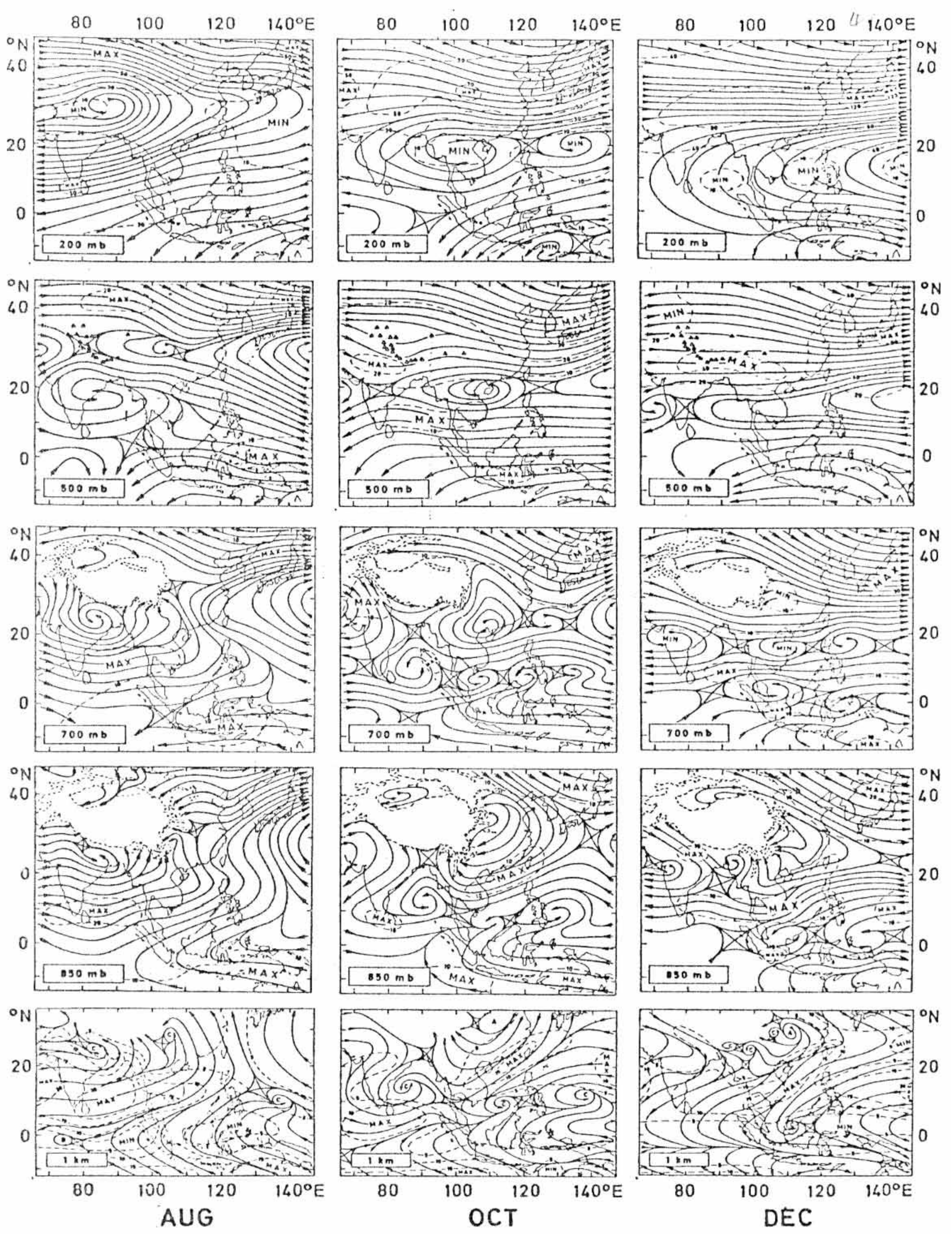

Fig. 3.25. continued

3.9 Air masses and fronts

The lower atmosphere tends to adopt the temperature of the ground or ocean with which it is in contact. Similarly, air over an ocean will adopt a higher humidity than that over deserts. Over some large and relatively uniform regions of the earth such as parts of the oceans, the poles and Siberia the temperature and humidity vary little over several thousands of square kilometres. The masses of air over these homogeneous "source regions" will attain characteristic temperature and humidity values - this leads to the concept of an air mass. It is possible to differentiate between warm (tropical) and cold (polar) air masses and between humid (maritime) or dry (continental) air masses. The four principal air masses are therefore called "continental polar, maritime polar, continental tropical and maritime tropical". When air masses move from their source region they are modified. For example, when continental polar air moves from land over warm oceans it is warmed and picks up moisture so changing its thermodynamic characteristics. It is therefore possible to define a whole range of air masses but little practical benefit is derived from doing this and so we shall confine our attention to the four principal classifications. Opposing air flows can bring warm and cold air masses together at a "front" - the name given to the narrow zone separating two air masses at the surface of the earth. Across a front there is a sharp change of temperature, which usually appears as a discontinuity on weather charts and station thermographs. When a front moves over the earth's surface under the influence of winds from the tropics so that warm air replaces cold air at its passage then we refer to the front as a "warm front", similarly, if cold air replaces warm then the front is called a "cold front". Fronts form in preferred areas in temperate and higher latitudes but they do not form in the tropics where, in general, the winds are relatively light and the temperature relatively uniform. However, cold fronts in the winter hemisphere do move into the tropics as subsiding polar air rushes equatorwards. For example, in the northern winter cold fronts lead invasions of polar continental air from Siberia into China and across the warm waters of the South China Sea. The polar air rapidly picks up heat and moisture so that at some latitude - usually about 15^{o}N but dependent on the speed of movement of the front and time of year - it can no longer be distinguished from the air which it replaces; the front then ceases to exist although, of course, the air may continue to move southwards and cross the equator. The process of modification is assisted by the shrinking in depth of the cold air as it spreads southward and the mixing caused by convection over the warm sea surface. The rapidity with which a front looses its nature is not necessarily dependent on the intensity of the polar outbreak or surge. Fronts ahead of major outbreaks sometimes dissappear faster than in lesser surges depending on the degree of subsidence and/or curvature of the trajectory.

During the winter, several cold fronts pass Hong Kong (22° 18'N) each month and some of them are detectable - as fronts - at Saigon (10° 47'N) but not at Singapore (1° 18'N). The long period absolute minimum temperatures at these stations are 0°C, 14°C and 19°C respectively and the average of the annual minima are 6°C, 17°C and 21°C. The latter figure for Singapore is only a little less than the 22°C average of the lowest temperatures recorded in each of the transition months, April and October, when of course, modified polar air of northern origin does not reach Singapore. Although the temperature and humidity differences between polar continental air and tropical maritime air usually disappear as the former moves southward across tropical waters, the two air masses can keep their respective velocities; in such cases what was once a front will retain an identity as a line of wind shear or "Shear line". The "ghosts" of old cold fronts may therefore penetrate to the equator - as shear lines - where they can sometimes be identified on satellite photographs as thin lines of cloud. Sometimes the associated convergence ceases and the shear line continues equatorward without cloud but,this may re-appear if convergence occurs again.

Attempts to detect differences in temperature or humidity between air masses reaching the equator in Oceanic areas from different source regions have, in the main, been unsuccessful. At equatorial stations diurnal and local variations of temperature and humidity are greater than the variations to be expected there between air masses, they therefore tend to conceal such air mass differences as might exist. John (1949) was unable to detect differences at any level in the air reaching Singapore from different source regions and Estoque (1952) found no true frontal difference in the Philippines between the north east monsoon air and that from the tropical Pacific. However, Gaigerov (1966) reported that a Russian research vessel sailing along the equator between 150°E and 180°E in May 1965 was able to detect small temperature differences up to heights ranging from 0.5 km to 2 km in 'tropical fronts'. It was claimed that air on the winter hemisphere side of the front was the cooler. However, the "fronts" were accompanied by rain and it is not clear how much of the reported temperature difference should be attributable to the cooling of air by the evaporation of raindrops. Even if small temperature changes can be attributed to air mass differences then fronts near the equator would have a structure greatly different from that found in the fronts of temperate latitudes.

Summarising, we may state that as cold fronts sweep southward over S.E. Asia and the West Pacific during the winter season each year, one or two will still be identifiable as fronts at 10°N or so. However, as they move further south they loose their frontal characteristics and proceed as shear-lines and if they retain any temperature/density discontinuities then they are so weak as to be difficult to measure. If any cloud or weather remains on the shear-line it will be found in a much narrower band than is associated with those fronts in higher latitudes.

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Extratropical cyclones

Just after the 1914-18 war the Norwegian school of meteorologists propounded the now classical frontal theory of the birth and development of cyclones. These cyclones - known as extratropical or frontal cyclones - form in temperate and higher latitudes on stationary or slow moving fronts which separate a cold dense polar air mass from a less dense warmer air mass. In response to divergence (removal of air) in the upper westerly wind flow, a wave develops in the front below with its crest in the cold air. Simultaneously, the pressure falls at the surface below the area of upper divergence. The lowest pressure is found near the crest of the frontal wave where a cyclonic wind circulation develops simultaneously - and a cyclone (synonymous with "depression") is born. It moves in an easterly direction under the influence of the upper polar westerly winds. Low level cyclonic winds spiral in towards the cyclone centre but the accumulation of air there, at low levels, lags behind the depletion aloft; the net eviction of air from the column of the atmosphere above the centre continues, thereby causing the pressure to fall further.

Initially the wind flow in the two air masses is parallel to the stationary front and is frequently in opposite directions, the cyclonic flow that develops around the new wave brings cold air equatorward behind the centre and takes warm air poleward ahead of the centre. The leading edge of the cold air forming the rear part of the frontal wave is therefore a cold front whereas the leading edge of the wave crest is a warm front. The area between the fronts is comprised entirely of tropical air and is called the "warm sector" (Fig. 3.2b). The developing cyclone is asymmetric, having fresh subsiding polar air behind the centre, tropical air in the warm sector and modified polar air returning poleward ahead, and poleward, of the centre. In the upper air the centre of lowest pressure is not directly over the surface centre but is found on a line which slopes with height towards the cold air. Initially, the circular isobars around the surface cyclone shrink with height until, at 3 km or so, only a wave in the upper westerly flow can be found.

As the cyclone moves, the cold front catches up the warm front. They join first near the crest of the wave and the junction then encroaches further into the warm sector. The union of the cold and warm fronts lifts the warm air which previously formed the warm sector between them and the resulting front is therefore known as an "occlusion" (Fig. 3.2c). This process of occlusion continues with the cyclone deepening i.e. central pressure falling and the wind circulation increasing, until all the warm

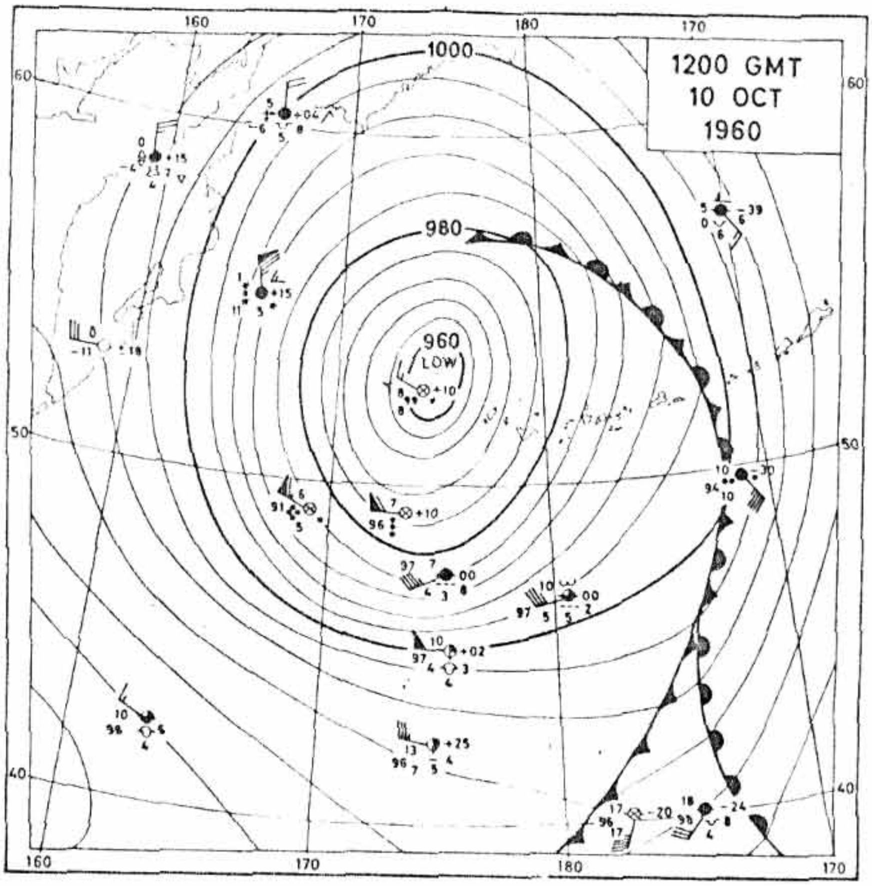


Fig. 3.26 Cold Front Warm Front Occlusion

An intense extratropical cyclone over the Aleutian Islands. The depression has a diameter of more than 1 600 km and gales are found as far as 1 000 km from the depression centre.

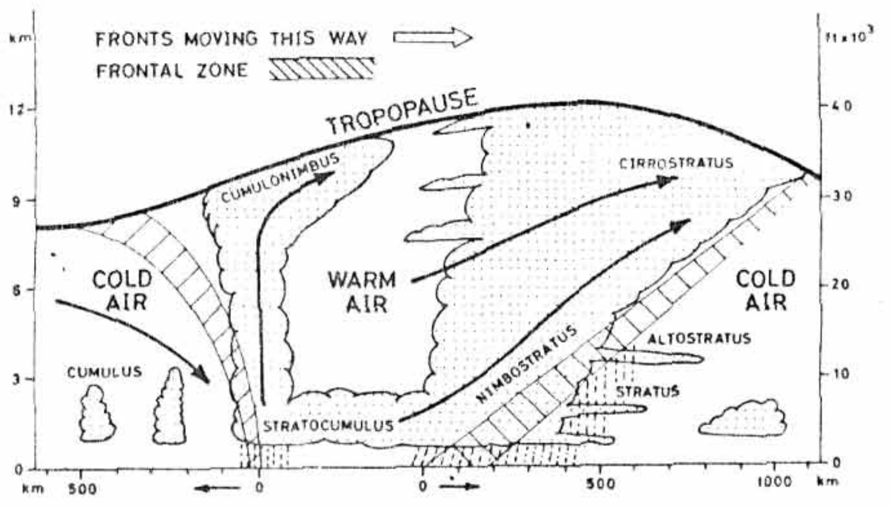


Fig. 3.27 A schematic vertical cross-section along an east-west line through a cold and warm front and the associated warm sector. A vertical section along the $40^{O}N$ parallel of latitude where it crosses the fronts shown in Figure 3.26 would show a similar structure.

sector air is lifted off the surface and replaced with colder air.

The weather map in Fig. 3.26 shows a deep depression (cyclone) over the North Pacific Ocean just before maturity. Figure 3.27 shows an idealised vertical section along latitude 40°N through the fronts and warm sector of the depression. In reality there are, of course, many variations on this generalised picture. It will be seen that the fronts slope with height towards the colder air. The cold front or rear part of the wave is quite steep in temperate latitudes, having typically a slope of approximately 1 in 50. The denser cold air underruns the less dense warm air forcing it upwards and producing a band of cloud and rain. When the cold front moves into the tropics the slope decreases to 1 in 200 or less and the cloudy area is then much broader. At a warm front, the tropical air in the rear rides up over the relatively denser colder air ahead. In temperate latitudes the slope of a warm front is about 1 in 150 and as the warm sector air glides up this slope a broad band of cloud and rain is formed. Satellite photographs of frontal depressions are shown in Fig. 4 and 13 .

The energy required to drive the cyclone winds against friction is obtained by converting the potential energy of the system into kinetic energy. This is achieved by the denser cold air moving down and under the warm air so lowering the centre of gravity of the system as a whole. However, when the warm sector is completely occluded the warm air will be aloft, the atmosphere will be stably stratified and the source of energy will be expended. The mature cyclone now begins to weaken or "fill in", with rising central pressures and decreasing surface winds. At this mature, occluded stage, the cyclone is more nearly symmetrical than at any other time in its life cycle, and a circulation of winds will often be found in the air above the surface circulation.

Typically, one to two days are required for each of the growing and decaying stages of the life cycle so that cyclones are usually active for 3 to 4 days. Each winter in high latitudes over the Atlantic and Pacific Oceans a few large cyclones grow to have diameters of 25 degrees of latitude i.e. 2,800 km or more and contain hurricane force winds over a relatively large area as shown in Fig. 3.26. The central pressure in these cyclones typically falls to about 950 mb (See also sect. 4.9.1).

The birth, energy source, structure and movement of extratropical cyclones are completely different from those characterisitic of tropical cyclones, in particular, fronts are not involved in the formation of tropical cyclones. The differences are discussed in section 4.9.1.

Equatorial Trough

We have ~~seen~~ noted that in the subtropics or "horse latitudes" - around latitude 30° - a belt of relatively high pressure is found at the surface in each hemisphere and between them, not far from the equator runs the "equatorial trough" of low pressure (Fig. ~~3.18~~ 3.3 and ~~3.22~~ 3.23). It contains the "doldrums" or areas of calm or light, variable winds so dreaded by early sailors. In the west Pacific and Indian Ocean regions the ~~monthly mean position of the~~ trough migrates north and south - lagging behind the seasonal march of the sun by about two months - and af~~fects the onset of~~ [illegible] the formation of typhoons and the weather generally at places in its path. In the east and central Pacific the ~~monthly mean position~~ latitudinal variation of the trough ~~moves through~~ amounts to only a few degrees during the year (Figs. 3.22 and 3.23) but in the west it moves further. The difference in behaviour is attributable to the greater seasonal change in temperature and pressure which takes place in the atmosphere over the Asiatic mainland compared to that over the tropical Pacific Ocean; in the former area "heat lows" or "monsoon lows" appear and form an extension of the equatorial trough. In general, around the world as a whole, the trough coincides with the belt of highest mean surface air temperatures or "thermal equator" and its mean position averaged over a year is found to be about 5°N rather than at the geographical equator.

The seasonal change in the distribution of pressure produces corresponding changes in winds, and, in particular, gives rise to the relatively persistent seasonal winds, the "monsoons" of South East Asia. In Fig. 3.~~18~~23 vector - mean winds are shown for places in the Pacific and Indian Oceans. Each wind is the resultant of a very large number of observations from islands and ships of many countries (   ) and represents the steady wind which would give rise to the observed net movement of air in each month. This net flow of air is depicted by streamlines in the figure where it will be seen that there is "confluence" or a flowing together of the winds, in the regions marked by heavy dashed lines. In the east and west these lines of confluence are near to the mean position of the equatorial pressure trough (cf. Figs. 3.22 and 3.~~18~~21) but this is not the case in the central Pacific where the mean lines of confluence, in both months, are found some distance from the equatorial trough.

The distribution of weather and winds along the trough is neither

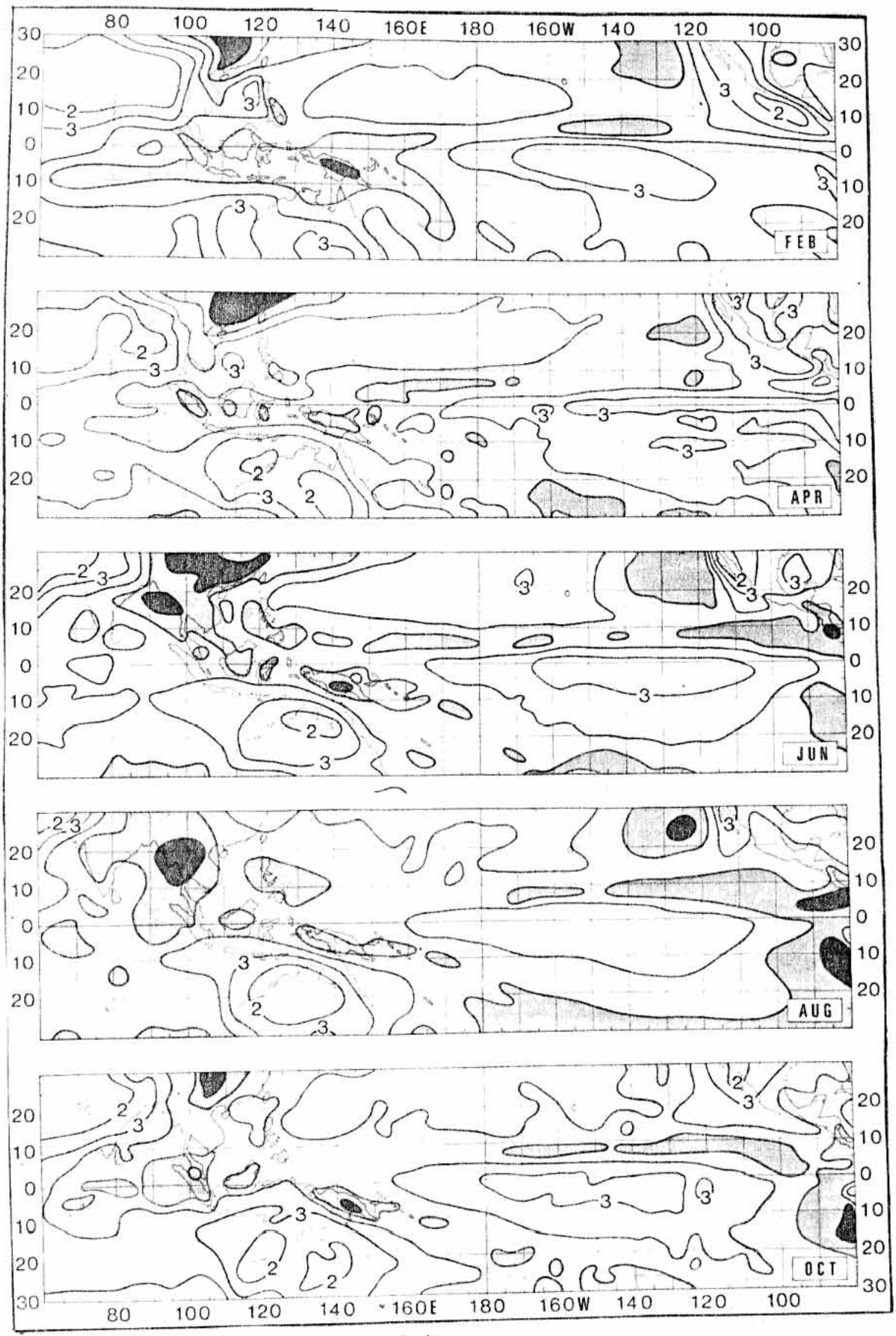


Fig. 3·17

Mean cloud cover during the three years 1965-1967 as determined from satellite pictures. The shaded areas show where, on average, more than 4,5 and 6 eights of the sky was covered with cloud - the higher values ~~being~~ are shown by darker shading. [Two bands of OUT cloud, one on each side of the equator, can be seen in the Eastern Pacific in April, at other times a single line of cloudiness is seen running east to west just north of the equator ~~areas~~ across the Pacific. In the west the cloud band ranges from just south of the equator to about 20°N. See Fig. 7

uniform in space nor in time. Between 180° and 120°W, for example, the northeast trade winds merge into the southeast trades to form one easterly airstream; this situation is relatively persistent and can be seen frequently on daily weather charts. Outside this area however, the airstreams from the northern and southern hemispheres often flow towards each other in opposing directions forming a line of wind shear which is found both in the mean and on individual days although the shear line is frequently manifest on daily charts in the form of small eddies or vortices. The opposing flows are found where air from the winter hemisphere crosses the equator and, because of the earths' rotation, turns towards the east to oppose the trades or air flow in the new equatorward hemisphere. Sadler (1967) has pointed out that the winds lose westerly on the the equator side of the equatorial trough where it is deflected about 100 or more from the equator. This statement is supported by the mean data shown in Fig. ~~3.18~~ 3.23. Note that in August the mean flow from the southern hemisphere turns to become a westerly flow in the northern hemisphere and so meets the North Pacific trade winds in opposition in both the east and west of the ocean but not in the central region. The steadiness of the trade winds between about 120°W and 180° accounts, to a large degree, for the resultant winds being stronger there (indicated by shading in Fig. 3.23) than in most places, furthermore, because of the steadiness of the trades the line of confluence moves little from its mean position. In the west however, the winds are, in general, less steady and on many days neither a line of wind shear nor a line of confluence, can be found.

.4 Near the equatorial trough there are usually areas covered by deep convection clouds and disturbed weather and, frequently, the convection is organized in broad lines although often broken into discrete cloudy areas now called "cloud clusters" (Williams 1970). These aligned cloud clusters are associated with low level convergence, and are therefore known as "convergence lines" or "convergence zones". A composite photograph of typical convergence zones in the equatorial trough, as seen from satellites, is shown in Fig. . The areas of disturbed weather with dense high cirrus clouds appear bright in these photographs and, typically, cover an area which could be enclosed within a rectangle having sides with a length of the order of 400 km. They have lives which vary from less than a day to about one week. We discuss cloud clusters on page because a few of them eventually become tropical cyclones.

.5 Since 1963 it has been possible to use satellite photographs to prepare monthly averages of cloud cover. Sadler (1968) has done this for two years and h~~e has~~ kindly supplied me with the mean cloudiness in one

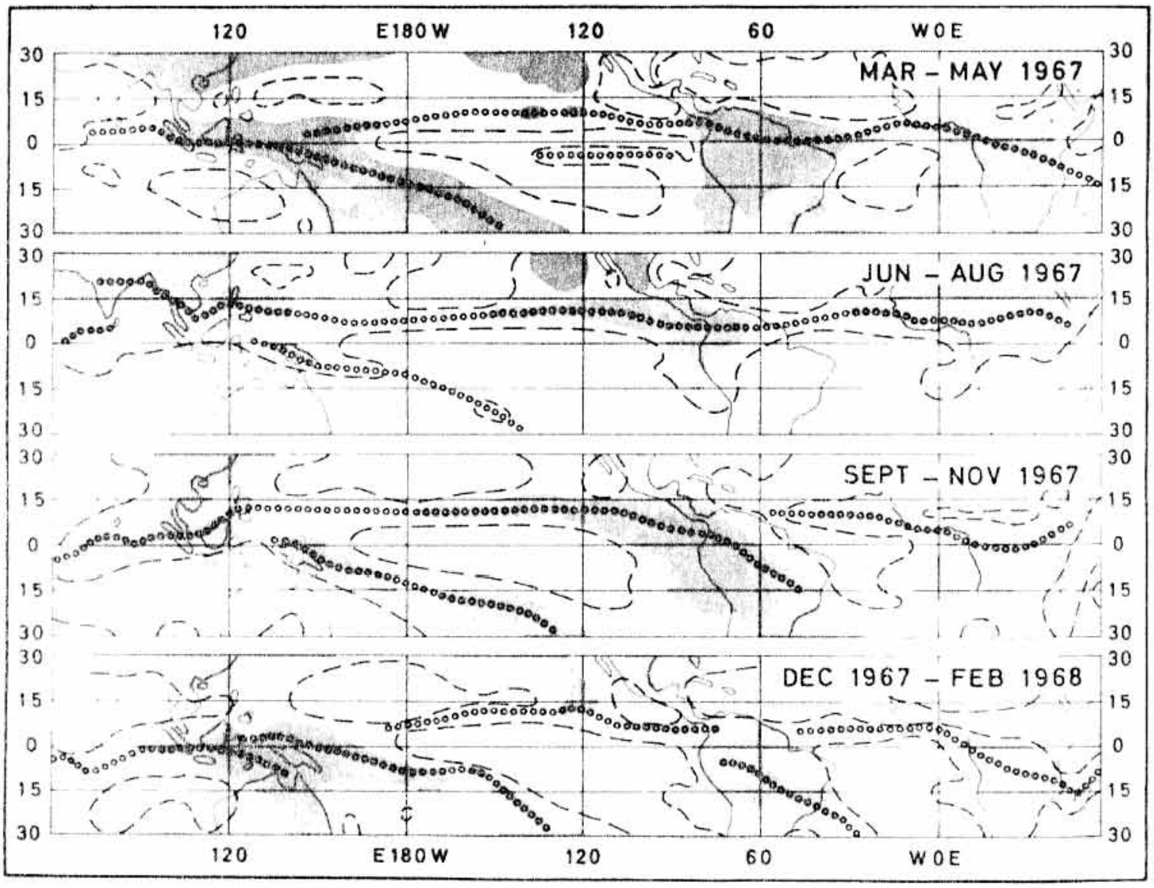
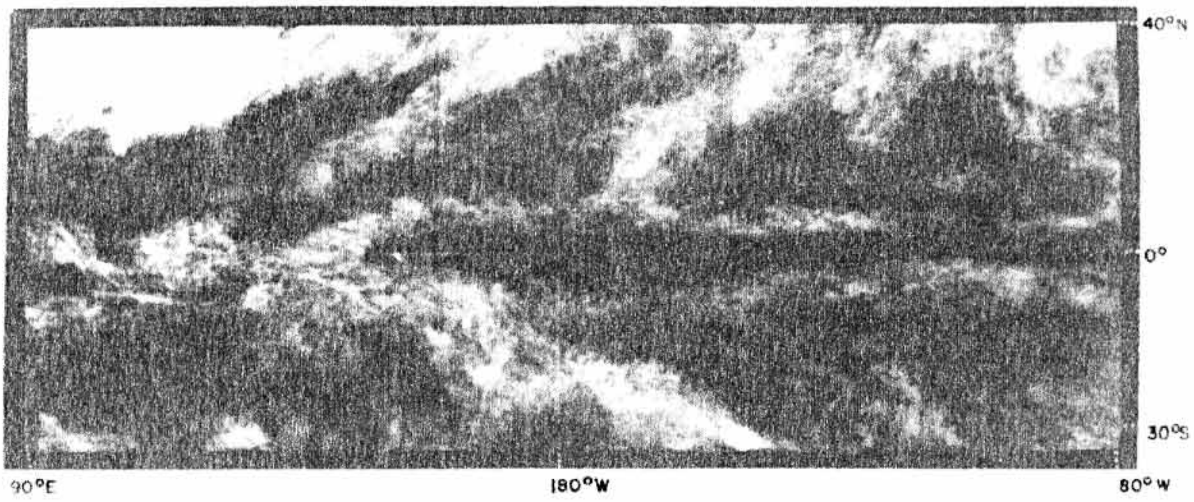


Fig. 3.~~20~~ 25 The seasonal distribution of the brightness of clouds as observed from satellites ESSA 3 and 5. Brightness values on the daily pictures were digitised and processed by computer to produce these maps. The dashed lines enclose areas of minimum brightness, areas of maximum brightness are shaded and the lines of circles indicate the major zonally oriented axis of maximum brightness (or cloudiness). (Prepared from an original diagram by Hubert et al (1969)).

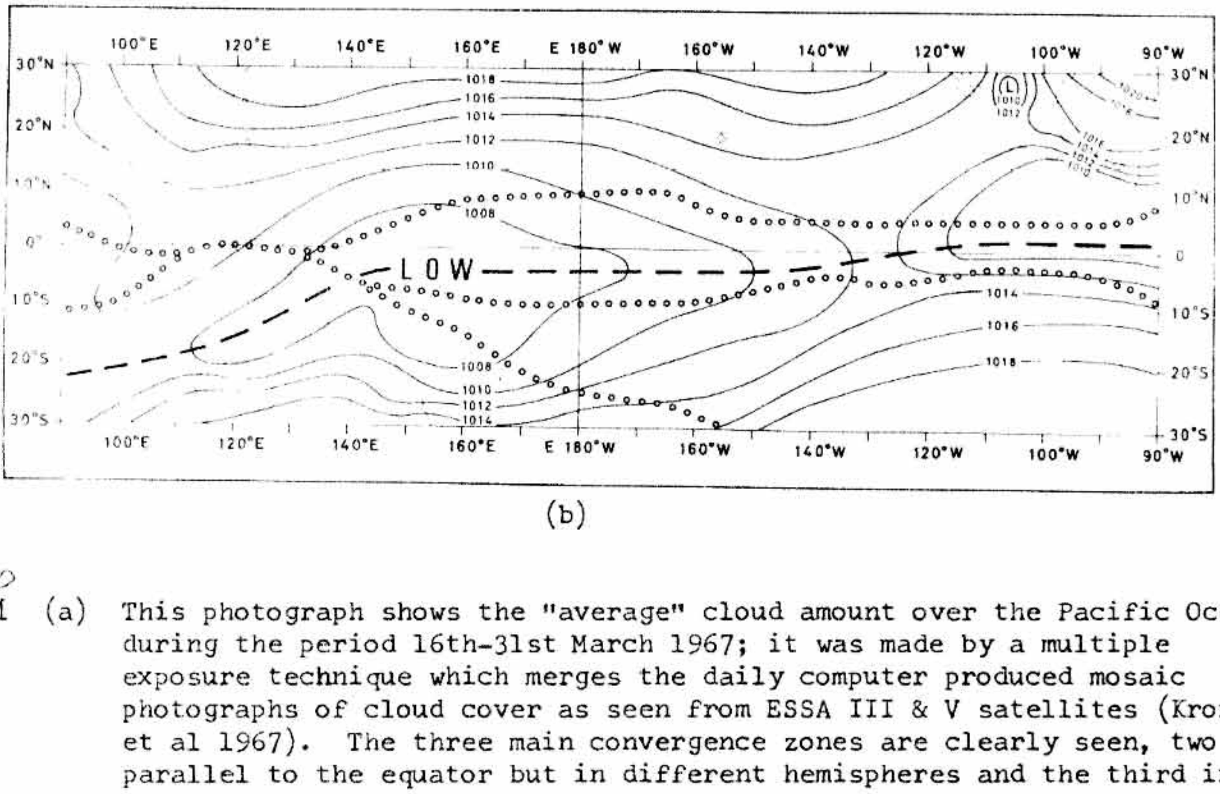
squares
degree ~~boxes~~ for the years 1965, 1966 and 1967. ~~and~~ I have used this data to prepare the charts shown in Fig. 3.19. They show clearly the mean position of the convergence zones which agree, in essentials, with those found by Hubert et al (1969) using quarterly averages of mean cloud brightness ~~for~~ in the years 1967/68 Fig. 3.~~20~~25. Three convergence zones over the Pacific Ocean and south-east Asia are shown on these charts. Firstly, there is a northern zone or band which is most marked during the northern summer. In the east, it remains relatively stationary during the year whilst in the west it ranges over about 25° of latitude. When the equatorial trough is south of the equator in the northern winter, the northern convergence zone is relatively weak and disappears completely in the west. Secondly, there is a persistent convergence zone in the southern hemisphere which moves very little, being anchored at its northwest extremity near New Guinea. This convergence line is most active in the southern summer when the equatorial trough is south of the equator (Fig. 3.2x ~~[illegible]~~
~~[illegible]~~
Lastly, a mean convergence zone appears south of the equator in the eastern and central parts of the Pacific Ocean during the spring months only. It is seen very well in Fig. 3.26 in which daily satellite photographs for the 15 day period from 16th-31st March 1967 have been averaged by a multiple exposure technique. This southern zone was clearly defined that year and this well publicised ~~known~~ photograph has created the impression that the double band of cloud is a permanent feature of the weather in the central Pacific; that this is not so can be seen both from the average cloud cover charts and from computer produced charts of mean cloud-brightness used to prepare Fig. 3.~~20~~25.

annual latitudinal variation
.6 The ~~movement~~ of the centre of the convergence bands at longitudes 140°E and 140°W is shown in Fig. 3.~~22~~17. It will be seen that the convergence zone lies about 5° to the north of the pressure trough at 140°W and somewhat less at 140°E. Examination of Figs. 3.~~17~~22 and 3.~~19~~24 show that the convergence zones remain near the trough in the east Pacific and Indian Ocean areas.

7 In summary, the equatorial trough is relatively stationary in the eastern Pacific but follows the movement of the sun over the west Pacific, southeast Asia and Indian Ocean. There is a confluence of airstreams - usually from opposite hemispheres - along lines in or near the trough. Sometimes and in some places these lines of confluence are convergent (i.e. produce a net accumulation of air) and so produce broad bands or areas of cloud and weather which are known as convergence lines or zones.



(a)



(b)

26
Fig. 3.~~21~~ (a) This photograph shows the "average" cloud amount over the Pacific Ocean during the period 16th-31st March 1967; it was made by a multiple exposure technique which merges the daily computer produced mosaic photographs of cloud cover as seen from ESSA III & V satellites (Kronfield et al 1967). The three main convergence zones are clearly seen, two parallel to the equator but in different hemispheres and the third in the Southern Hemisphere running from northwest to southeast across the dateline.

(b) The monthly mean sea level pressure from March 1967. The location of the convergence zones shown in (a) are indicated by open circles.

In most cases the surface wind velocities are such that convergence takes place some distance to one side of the ~~lines of confluence~~ trough. Although the equatorial trough is associated with lines of confluence, lines or zones of convergence and shear lines they need none of them be coincident except when the pressure trough lies along the equator (Ramage 1970). Three convergence zones appear on charts of monthly mean cloud cover, the zones are relatively inactive or disappear altogether at times when they are distant from the equatorial pressure trough and, apart from the western part of the northern zone, they do not move far from the equator during the year.

3.10.1 The Intertropical Convergence Zone or ITCZ

The convergence zones in the equatorial trough were early observed by mariners, and later by aviators, to be ~~relatively narrow~~ bands of convection cloud which meteorologists considered to mark the zone towards which flowed the trade wind systems of the northern and southern hemispheres. In particular, the northern convergence zone, ~~discussed in the previous section,~~ was given the name "Intertropical Convergence Zone or ITCZ".

It will immediately be noted that two interpretations of the term are possible because the line defined by the wind flows and that defined by the cloud are not, in general, coincident. However, in addition, several other definitions have been given and the term has frequently been used where the causal connotations were not relevant. For these and other reasons the term ITCZ has fallen into disrepute amongst many tropical meteorologists. I had hoped to avoid using it but, since it is still found in many texts and papers, the reader will need some account of its history and usage.

The very large and confusing literature on the ITCZ contains many conflicting statements some of which have been detailed by Palmer (1951). The confusion has arisen from the inadequacy of data in earlier years, from regional differences in the method of formation of the ITCZ and, particularly, from the use of different properties to define the location of the zone. Pressure airstreams, rainfall, cloudiness and steadiness of the wind have all been used for this purpose, however, the positions determined by the different methods ~~will~~ are not, in general, ~~be~~ coincident. In addition, because the connotations

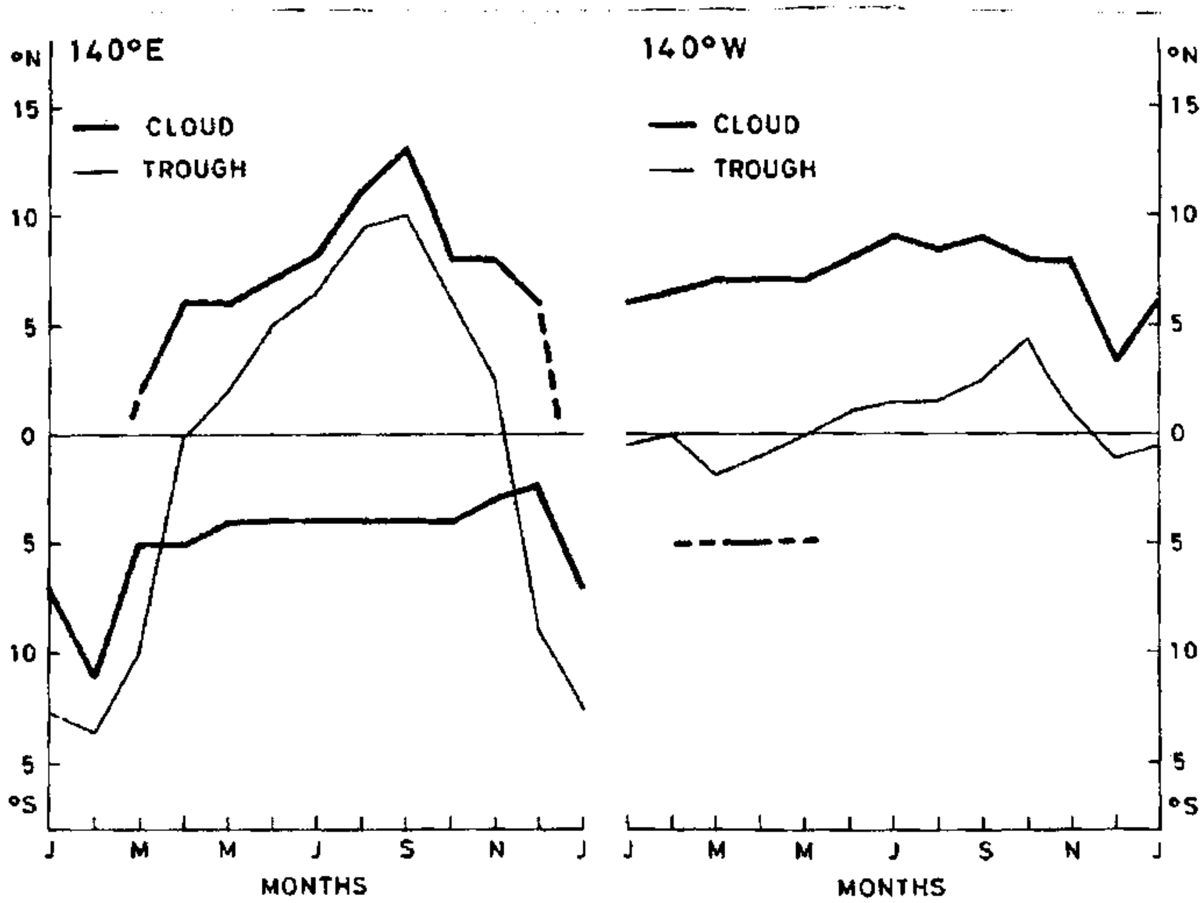


Fig. 3.~~22~~ 27 The seasonal latitude of the equatorial trough at 140°E and 140°W is shown by the thin lines whilst the thick line shows the latitude of the near equatorial convergence zones at the same meridians. The position of the convergence zones was determined from the axes of maximum cloud cover shown on Sadler's maps for 1965 and 1966 (Sadler 1967).

Put line on top of diagram

of the term "intertropical convergence zone" are too specific to be of general application to the near-equatorial convergence zones found around the world, new terms have been coined in the hope that they would be more widely acceptable but they too have not been sufficiently general and have served only to increase the confusion.

Historically, Brooks and Braby (1921) in their paper with the descriptive title "The Clash of the Trades in the Pacific", [8], used the term "equatorial front" to describe the line along which the trade winds meet, Fig. 3.23 The success of the Norwegian air mass and frontal theory in temperate latitudes, together with the realization that airstreams from different hemispheres meet in the equatorial trough region, led Bjerknes (1930) to coin the term "intertropical front" to describe the same phenomenon. By the mid-mineteen forties it was generally agreed that the intertropical front was not a front in the accepted sense because there were no significant differences in the temperature or density of the airstreams involved ~~(p.)~~. The term was therefore changed to "intertropical convergence zone or ITCZ". However, there is still too much implied in this name because the airstreams on either side of the trough need not necessarily have come directly from different hemispheres, for example, westerlies from India often meet the North Pacific Trades ~~Fig. 3.18(b)~~. To avoid this objection the term "equatorial convergence zone" was introduced. However, airstreams are often confluent (flow towards each other) without being convergent (accumulating mass) so that "equatorial convergence" is not always an appropriate term and is especially inappropriate when the area concerned is found over 1,000 km from the equator. For example, in the mean, the line of confluence between the South-West Monsoon and the North Pacific Trades runs north of Luzon in August (Fig. 3.21b) and lies along the equatorial pressure trough, Fig. 3.22, but it is not usually a zone of convergence because it is seldom associated with cloud or weather (Fig. 3.24) although there is adequate water vapour for cloud to form if there were low level convergence.

New terms have been coined and are still being coined in attempts to overcome these deficiencies. Recently, for example, Palmen & Newton (1969) have referred to the "trade confluence or TC" whereas the Regional Tropical Analysis Centre at Miami U.S.A. has adopted the term "Intertropical

Possibly not so in Atlantic. However, the term does not seem to have been generally adopted. [illegible]

Confluence or ITC" defined by Simpson (1968) as, "a nearly continuous fluence line (usually confluent) representing the principle asymptote+ of the Equatorial Trough". The ITC is here defined with reference to the circulation or wind patterns which, in ~~general, will~~ need not be coincident with either the equatorial trough or the associated convergence zones, the latter being called the ITCZ by the Miami Centre. No terminology for cloud bands usually found to coincide with asymptote.

In view of the foregoing discussion and the obvious need to avoid putting too much into a name I follow the practice recommended by Riehl (1954) in his classic book on tropical meteorology, and use the term "equatorial trough". However, because the equator is a precisely defined region of the earth, and because the trough migrates some distance from this line - especially over Asia in the summer and because there are sometimes two troughs with westerlies between them ~~- purists may prefer to follow~~ Ramage (1970) ~~and refer to a~~ prefers the term "near equatorial trough". Either of these terms is satisfactory, they are specific, and do not imply a causal connotation; they allow that in or near the trough may be found "lines of confluence", "shear lines" and "convergence zones or lines".

3.10.2 ~~Some~~ Characteristics of Near Equatorial Convergence Zones

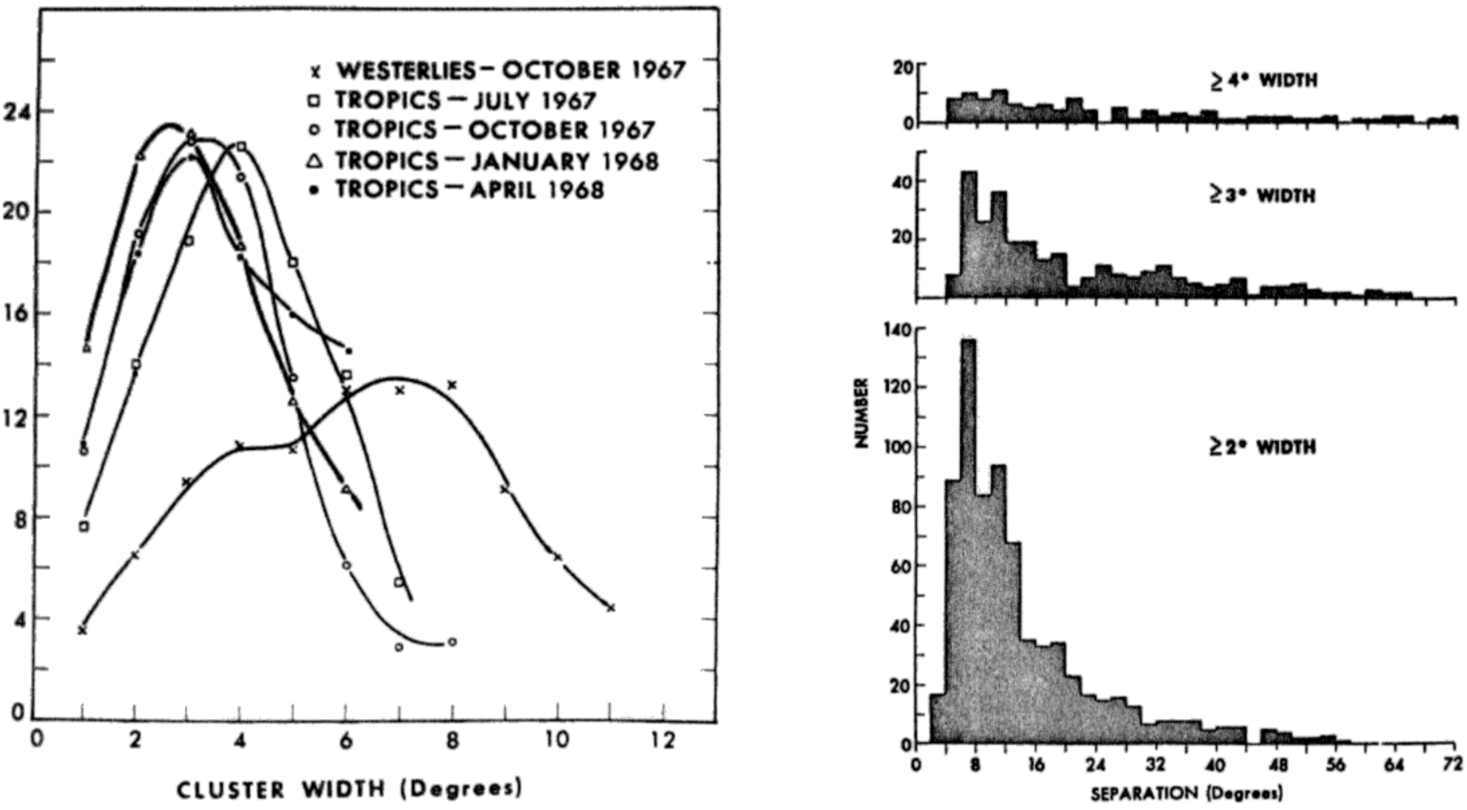
The lines of convergence that are found in or near the equatorial trough are comprised mainly of large cumulo-nimbus clouds whose spreading anvils combine to form a cirrus canopy. At low levels air converges towards the cumulo-nimbus towers, rises and flows outward mainly at the level of cirrus clouds. The converging air near the surface produces smaller cumulus clouds around the main mass of cumulo-nimbus towers, and sheets of alto-stratus clouds form at mid-levels as the humidity rises there from water vapour carried up by the cumulo-nimbus clouds. Although long continuous cloud bands are found (Fig.) the convergence zones are more frequently broken into cloud clusters (Fig.) whose east-west regimentation define the convergence line or zone. The following discussion relates mainly to convergence zones as they are found over the tropical Pacific Ocean.

+ Asymptote is a line along which the confluent airstreams merge.

Hayden (1970) has shown that cloud clusters in near-equatorial convergence zones over the Pacific have widths which are predominantly in the range 275 to 450 km as shown in Fig. 3.~~26~~28a with a tendency for the larger sizes to occur in the summer. The frequency of occurrence of various distances between the clusters is shown in Fig. 3.~~24~~28b and Hayden points out that in all three distributions the spacing of 6° - 8° (780 km) and 10° - 12° (1200 km) appear as maxima. He considers the larger spacing to be the principal synoptic scale in tropical convergence zones with the smaller scale being due to interaction with faster moving secondary disturbances. The number of cloud clusters which formed in selected months in 1967 is shown in Fig 3.~~25~~29

Some of these cloud clusters ~~in~~ extend well into the easterly trades ~~have an extended life~~ and propagate ~~to the west, such~~ westward; such moving disturbances have been studied by Chang (1970) using a novel method. Daily satellite photographs of the tropical north Pacific were cut into strips of width equal to 5° of latitude. When laid ~~next~~ to each other~~,~~ in date order, as shown in Fig 3.~~26~~30 a persistent, stationary cloud cluster would appear at a constant longitude on all strips; persistent, moving clusters would change longitude and therefore lie along sloping lines. From the two ~~such~~ composite photographs ~~are shown~~ in Fig. 3.~~26~~30 ~~from which~~ it is clear that there are many moving persistent clusters. Chang found that they ~~clusters~~ tend to move westward at about 9 m/s or 7° latitude per day and are spaced at distances between 2000 - 5000 km (20° - 45° latitude). Clusters therefore pass a given longitude most frequently at intervals of 3 to 6 days. The speed and distance between clusters is therefore of the same order as that of waves in the easterly trades (p.) and it is most probable that the two are associated and that the clusters move with the speed of waves rather than being simply advected by the prevailing mean wind.

Not all cloud clusters persist. Braak (1821 - 29) long ago pointed out that in equatorial convergence zones weather often developed and decayed quite rapidly in situ. Satellite photographs now enable us to see that, indeed, some cloud clusters form and die within 12 hours or so whilst others persist and move for several days and yet others move northward out of the convergence zone to become tropical cyclones. What is true of individual cloud clusters is also true of equatorial convergence zones as a whole, that is, they too can quickly decay in one location and be replaced by new ones at some distance to the north or south. This process gives rise to a discontinuous



(a) (b)

Fig.3.~~23~~ 28 (a) Relative dominance of clusters of different sizes in the near equatorial convergence zones for individual months. Relative dominance is obtained by weighting the frequency of a cluster size by its area and dividing by the total area of all cloud clusters. The analysis was restricted to the brightest band of width 10° latitude. A temperate latitude sample is included for comparison.

(b) Frequency distributions of cloud cluster separation distances in the near equatorial convergence zone for a four month period. Smaller clusters have been progressively excluded in the upper histograms. (From Hayden 1970).

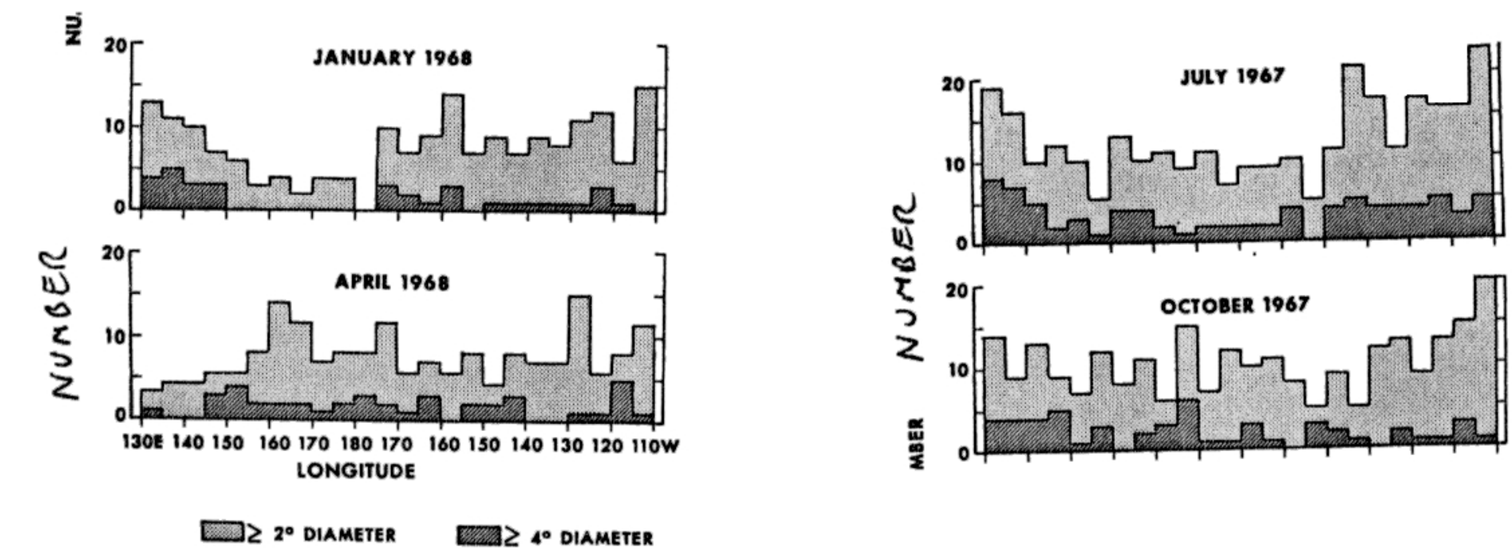


Fig. 3.~~24~~ 29 Zonal distributions of cloud clusters by month for the latitude band 0°-15°N. Note that more big clusters are found in the west Pacific than in the east. (From Hayden 1970).

movement of the zones which is completely different to the behaviour of fronts which are ~~usually~~ advected progressively from one place to another. This apparent jump in the position of convergence zones puzzled many meteorologists in the nineteen thirties and forties at a time when they were trying to apply temperate latitude frontal theory to convergence zones.

~~The latitudinal position of~~ The convergence zone migrates to the north or south mainly in response to the relative intensities of the low level circulations on its flanks. However, the mechanisms for the day to day variations in position are not yet fully understood ~~but~~ time lapse films of satellite photographs show marked changes in activity and position of the zones when deep troughs in the upper westerlies and associated cold surges pass by. In this respect, the activity and location of convergence zones is responsive to strong developments originating in temperate latitudes (BELL 1969). Changes in the subtropical ridge and jet stream also affect the position of the zones as do sea temperatures (Ramage 1970).

From a study of mean monthly maps of cloud cover for a three year period ~~(~~3.15)

Sadler (1970) has shown that the equatorial convergence zones do not seasonally move across ~~cross~~ the equator; that is, the mean convergence zones shown in Fig. 3.25 do not cross the equator in their north and south migration but rather weaken or dissipate in the hemisphere going into winter, and form or intensify in the hemisphere entering summer. The equator therefore lies in an area of minimum cloudiness throughout the year.

Sadler (1963) pointed out that in the eastern Pacific the main convergence band usually lies in westerly winds about 150 - 300 km south of the pressure trough and that a weaker secondary band of cloud appears a similar distance north of the trough in the easterlies. I have shown (Bell 1969) that ~~presented (1969)~~ ~~rainfall~~ ~~observations that indicate that~~ summer monsoon troughs near the China coast ~~have~~ two rainfall maxima, one to the south and one to the north of the trough. Tabamo (1970) also found that troughs near the Philippines had a similar pattern of rainfall with more rain falling in the westerly winds on the south side of the trough than in the easterlies to the north.

Finally, there is a popular misconception that every flight across the equatorial trough will encounter rough weather in the "ITCZ" which will be found close to its normal climatological position. In fact, if we consider the

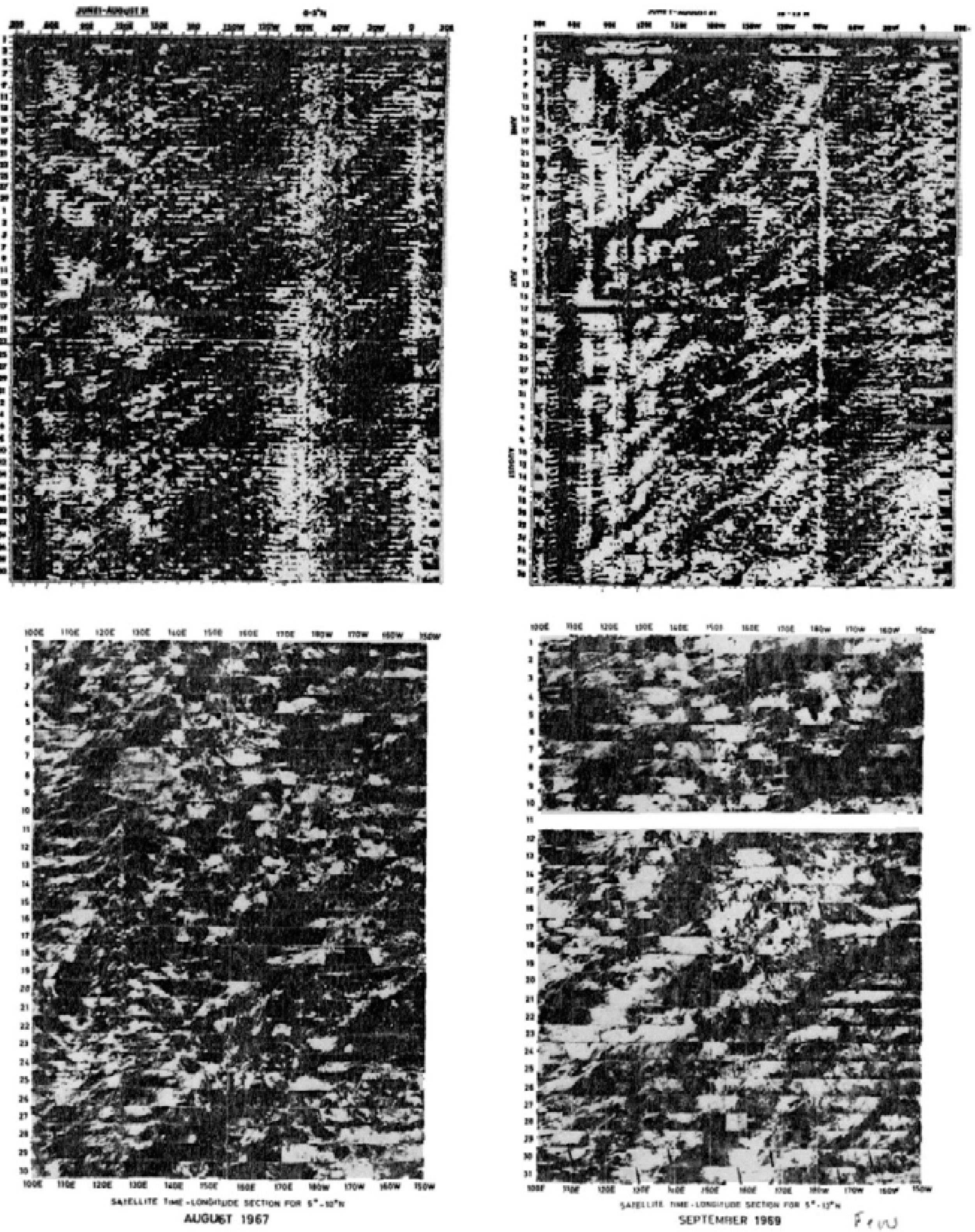


Fig. 3.~~25~~30 Strips of satellite photographs showing the cloud cover around the world, (a) between 0 -5°N and (b) between 5 -10°N, arranged in date order for June, July and August 1967 (*From Wallace 1970*). *Note that some equatorial cloud clusters* move eastward across the Indian Ocean in June & July - top left in (a) - as far as 150°E. *Whereas, cloud clusters originating over Africa - extreme right of* pictures - can be seen to progress westward across the Atlantic and occasionally cross America to join the eastward march of ~~other~~ clusters across the Pacific as far as Indo China and sometimes beyond. Pictures (c) and (d) cover the Pacific and the Far East only and are *for August 1967* and September 1969 *which* were months when the number of tropical cyclones which ~~to~~ formed was greater (10) and less (6) than normal respectively.

"ITCZ", to be defined by convergence as indicated by cloud~~s~~ bands then it will frequently be found well away from the mean positions shown in Fig. 3.~~10~~ 20; this is especially true in the western Pacific. On some days a convergence zone will be entirely missing in some regions and even when such a zone is well marked it will frequently be split into cloud clusters with fine weather between. Convergence zones are not always traceable on a day to day basis but the equatorial trough with which they are associated can always be found as a near equatorial pressure minimum between the sub-tropical anticyclones of the northern and southern hemispheres.

3.10.3 <u>Definitions.</u> The following definitions may be of use to the reader.

1) "Equatorial trough" or "Near Equatorial Trough" – ~~the~~ a region of low pressure found between the sub-tropical high pressure belts of each hemisphere. At times there may be more than one trough with westerlies between them. ~~The~~ The trough may contain fine weather or one or more of the following systems.

2) "Convergence zone" and "convergence line" – a zone (or line) of low level convergence which, over the tropical oceans, is usually indicated by considerable cloudiness and disturbed weather. The words "equatorial" and "tropical" are used as ajectives of location and in particular, the term "The equatorial convergence zone" is often used to describe the more persistent northern convergence zone in the eastern Pacific.

3) "Equatorial shear line" – a line separating airstreams having markedly different speeds or directions – usually equatorial westerlies and easterly trades – and lying within ~~or near to~~ the equatorial trough. Eddies or vortices occur along shear lines. Usually westerly winds are nearest the equator so that the eddies will be cyclonic in whichever hemisphere they form and when they are displaced poleward, as occurs in the summer hemisphere, the eddies are accompanied by a fall of pressure at their centres and disturbed weather. Eddies close to the equator have no associated low pressure centre and are usually inactive, that is, without low level convergence and bad weather. Except right on the equator a shear line will be accompanied by a trough.

4) "A line of confluence" - a line along which two airstreams merge without velocity convergence. In other words, it is a line towards which two airstreams converge without causing local accumulation of air; this condition requires that air be carried away horizontally from a region as fast as it is brought in; this is usually achieved by the wind increasing in speed in the downwind direction.

5) "A line of diffluence" - a line from which airstreams diverge. (If found at low levels) It is usually accompanied by fine weather.

3.11 Cloud clusters

Wallace (1970) presented photographs which show the progression of cloud clusters around the globe between various latitudes. Fig. 3.25 (a) and (b) show how cloud clusters moved in summer 1967 between 10-15°N and 0-5°N respectively. It will be noted that the westward moving cloud clusters- that is those that slope from top right to bottom left - are abundant and that some of those found over the Pacific have their origins over Africa which lies on the extreme right of the photographs. Because typhoons form from cloud clusters it follows that some typhoons ~~may be said to~~ could have originated in Africa.

Wallace's pictures for 0-5°N in May show ~~well developed~~ clusters over the equatorial Indian Ocean marching eastward and reaching 150°E by the end of the month, in weaker form, they continue eastward in June, right across the Pacific, or June ~~in June~~ as can be seen in the top ~~15 days~~ centre of Fig. 3.25 (a). A strong cluster ~~is seen~~ was moving eastward over the Indian Ocean at the same time as can be seen in the top left of the photograph. By the end of June westward moving disturbances ~~take over~~ replace the eastward moving clusters in both latitude bands.

1) Paul Chang first hype.
2) Difference between clusters in 1967 and in centuries

Waves

Westerlies –

Long waves relation to storm
recurving etc
global and short.
Satellite pix
Relationship to cyclones

Easterlies:

Inverted V – normal pix like an [illegible] but
more persistent cloud cluster.
[illegible] its virtue
Radar pix importance of Cb over
sea.

The satellite view of an easterly wave is more like a large persistent cloud cluster - covering an area similar to that indicated in Fig. 3.12 (1a). Radar pictures show a more banded structure Fig. 3.12(2) corresponding to the rainbands.

Riehl (1954) indicated that rain forms at and just behind an easterly wave trough axis when the wind blows through the wave and at or just in front of the trough axis when the wave moves faster than the wind. This simple concept is more complicated when the wave axis slopes with ehight and synoptic considerations based on the concent of upper divergence over lower convergence producing vapouous convection with the opposite structure producing fine weather. Indeed because of the varied nature of easterly waves different nomenclature for the various variants have been proposed to differentiate them from the classical easterly wave described by Riehl. However, this is not practical and waves in the easterlies have to be related to the speed and sheer of the basic current to be properly diagnosed.

Westerly waves

Although a feature of the temperature latitude these waves have much to do with the formation and movement of tropical cyclones in addition, in the winter months, they can have such large amplitudes as to plunge as far south as the equator. They differ from easterly waves in these amplitude and wavelength and the fact that they have a baraclimic structure. There are most frequently four waves around the N.H. but this can change to as few as two or as many as . A southward plunging westerly wave can enhance the amplitude of any easterly wave that should pass its longitude. An interaction of importance to forecasters. Although the forecasting of the development and moment of waves is then westerlies is of great importance to the tropical cyclone forecast problem, it is a subject that is covered in normal meteorological texts and are in which numerical methods have achieved considerable success over periods of 3 days or more at latitude above about 28^{o}N or south.

3.12 Easterly Waves

The low level and mid-tropospheric easterly flow around the equatorial flank of the sub-tropical high pressure belts is, relatively steady by comparison with the westerlies. However, wave-like pertubations in the flow occur and propagate westward in association with small changes in surface pressure (Dunn 1940) and increased cloud and rain. These disturbances were described by Riehl (1954) and are called "easterly waves". They are more frequent in the Atlantic than in the western Pacific. However, the structure of these waves changes as they move from the east to the west Pacific (Reed & Recker 1971) and there are differences between Pacific and Atlantic waves (Reed et al 1977). The variations in structure and occurrence of these mid-tropospheric waves is now well documented and, in particular, the Global Atmospheric Tropical Experiment (GATE 1974) permitted those waves which propagate across the Atlantic from their source region over Africa to be very well defined. Reed et al (1977) composited observations made in waves which crossed the GATE area during a 28-day period. The compositing was performed relative to the wave trough at 700 mb i.e. to the longitude of vorticity maximum at 700 mb. The whole wave was divided into 8 longitudinal categories as shown in Fig. 3.12(1) a. In the mean, the latitude of the vorticity maximum was found at 11^{o}N; the wavelength $\sim$2 500 km ($\sim 22^{o}$ lat.); the speed $\sim$8 m/s and the average period $\sim$3.5 days. The waves form in the period June-September over the African continent as a result of dynamic instability in the mean easterly current. The energy sources for growing waves comes from both the kinetic energy of the mean current and the potential energy associated with the temperature difference between hot desert air to the north and the cooler south westerly monsoon current to the south. This difference giving rise to an easterly thermal wind and an east-wind jet of about 10 m/s near 600 mb. Shear occurs in both the vertical and along a meridian (Fig. 3.12(1)(a)). The disturbance is most intense at 650 mb, being cold core below and warm core above. The main area of convective cloud and rain occurs just to the west of the trough and just south of the vorticity maximum as is shown in Fig. 3.12(1)(c). The strongest upward motion (5 mb/h) occurs at the 700 mb level just ahead of the trough.

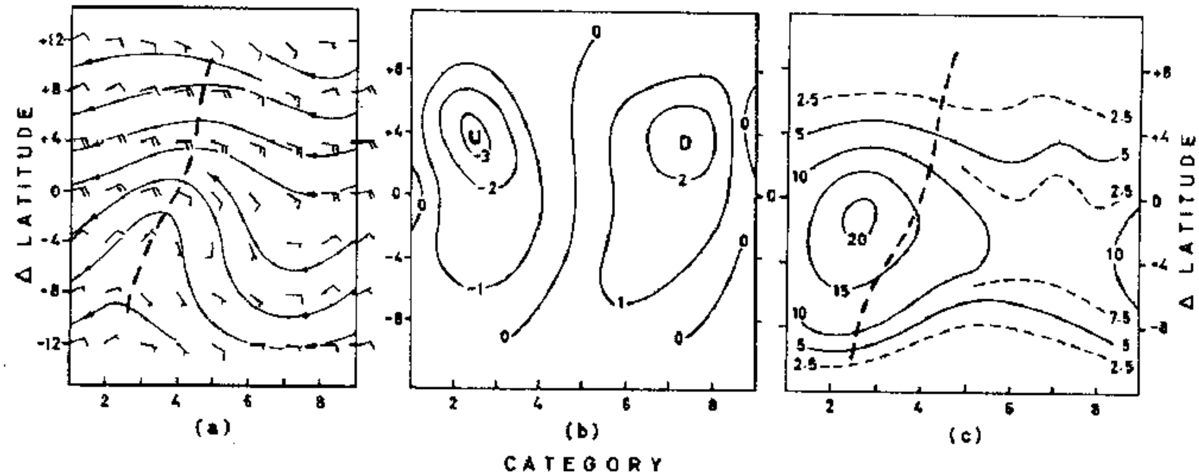


Fig. 3.12(1)(a). A composite streamfield at 700 mb of tropical easterly waves in the eastern Atlantic. The latitude relative to the vorticity maximum is specified on the ordinate. The mean latitude for the vorticity maximum is ~11°N; the mean wavelength is ~2 500 km (equal to 8 wave categories). A full barb on the wind vector indicates 5 m/s. The dashed line is the trough axis of maximum vorticity. (b) The composite vertical motion field (in mb/h) at 850 mb. (c) Average precipitation rate (in mm/d) from shipboard measurements. The trough axis at 700 mb is indicated by the dashed line. Cloud cover is given approximately by a similar pattern with the numbers doubled to give percent cover. (Adapted from Reed et al 1977)

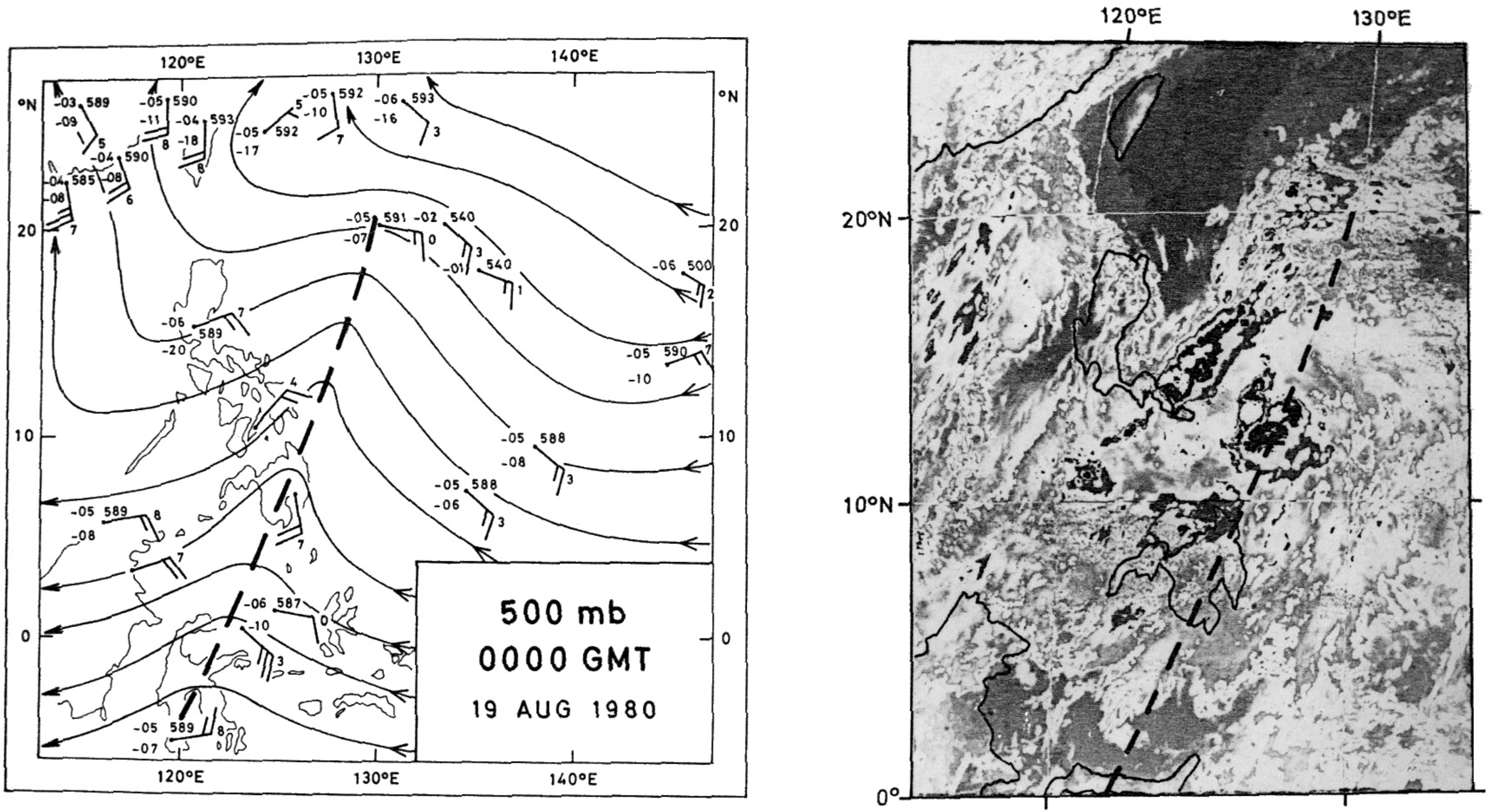


Fig. 3.12(2). An easterly wave at 500 mb (left). The distribution of deep convection clouds about the wave axis is shown in the enhanced infra-red image, for the same time, from the GMS satellite. The white cores inside the black areas show high cloud tops with temperatures of about -80°C. The warm Kurishio current can be seen as the darker coloured sea to the east of Taiwan.

Present studies support the barotropic-baroclinic instability growth mechanism for these waves and the CISK mechanism is not considered necessary to sustain them although the low level vorticity field is generally related to the vertical motion and cumulus convection. In contrast, Nitta (1970) and Wallace (1971) infer that latent heating is an important factor in maintaining easterly waves over the western Pacific. In this region the available potential energy generated by the wave cumulus clouds acts directly to maintain their kinetic energy by baroclinic energy transformations or, in other words, the heating due to convection generates baroclinic conditions which sustain the wave. A secondary but important source of wave energy comes from middle latitudes.

The structure of the waves undergoes a systematic change as they travel ~~from the~~ across eastermost ~~to the~~ wester~~most islands of~~ the Pacific. In the east The wave axis ~~initially~~ tilts eastward with height, as in the original Riehl model, becomes vertical in the central region and slopes to the west ~~over islands~~ in the western Pacific (Reed and Recker 1971). Precipitation and cloud maxima tend to occur to the east of the trough axis at eastern stations (cf GATE area waves Fig. 3.12(1) c) and to the west of the axis at western stations (135^{o}E). The different characteristics of the waves in different regions is one of many cases in which regional differences in wind, temperature and moisture fields lead to variations in the characteristics of disturbances in the tropics. The changes are mostly due to the variation with lo n gitude of the vertical shear of the basic easterly current and to its speed. There also seem to be differences between the waves over land and those over the ocean (Reed et al (1977)) and sea surface temperature has also been shown to affect the structure of the waves and to be a cause of the year to year variation in both their structure and frequency (Chang & Miller 1977). (A)

In the China Sea region Conover (1973) found that easterly waves moved through the southwest monsoon and were a major cause of weather changes there. The waves ~~were~~ are basically the same as those in the western Pacific but they are not readily perceived by ordinary analysis as they are obscured by the monsoon. Their wavelength and period are longer - 3 100 km and 5.5 d - and their speed slower (6.5 m/s) than the waves further east. Also, the maximum relative vorticity occurs at higher levels (700-500 mb) implying a lesser dependence on convergence

(A) Riehl (1954) pointed out that the distribution of ~~weather~~ convergence and rain would differ according to whether the wind blows through the wave or the wave moves faster than the wind. Low-level conve in the boundary layer. The rising motion and cloudiness fed into the vorticity centre from the south and southwest, to the north of the centre subsidence and decreased cloudiness occurs. The movement westward of enhanced convection as the waves penetrate the southwest monsoon is shown dramatically on the composite of strips of satellite photographs such as the one shown in Fig. 3.

The easterly wave model described by Riehl provoked considerable controversy in the 1960's. ~~because it had~~ The model had certainly been grossly overworked, disturbed-weather areas in the tropics were incorrectly and frequently interpreted within the frame work of the easterly-wave model. Geographical and inter-annual differences in both the nature and frequency of easterly waves added further fuel to the controversy. However, ~~from the foregoing~~ it is now clear that the classical easterly wave is not very frequent in the ~~eastern~~ western Pacific and is but one of a hierarchy of tropical perturbations which range from cloud clusters with weak relative vorticity maxima ~~through~~ to perturbations with a larger vorticity maxima due to zonal shear or wave-like flow ~~through~~ to disturbances in which a vortical arrangements of clouds can be detected ~~to~~, followed finally by the intense typhoon. The formation, structure and intensity of each depending on the broadscale flow and other environmental conditions.

An inverted V formation of clouds has been observed in the Atlantic and associated with the low level flow in easterly waves (Frank 1969). However, using improved satellite surveillance, Fett et al (1974) attribute the pattern to a cyclonic relative vorticity maxima north of the near equatorial convergence zone. ~~This draws air~~ The inflow to this centre causes across the equater ~~during~~ a drawing the initially linear cloud band into an inverted V. The western ~~area~~ leg of the V joins the cloud vortex leaving a break in the convergence zone ~~broken~~.

Easterly waves in their various forms are important in local forecasting in the western Pacific area as they often interrupt prevailing spells of fine weather on the south and southwest flanks of the Pacific ridge. Weak waves showing nothing more than cumulus clouds can approach the South China coast and combine with surface heating to release vigorous convection and thunderstorms. In this area, fine weather usually precedes the trough and indeed the abnormally clear skies and good visibility in the subsident area is usually a sign of an approaching wave - an useful indicator if heavy clouds have not already developed. Tropical cyclones frequently develop from easterly waves in the north Atlantic region but such development is relatively rare in the western Pacific. However, easterly waves occasionally occur in the latter area in the high troposphere (200-300 mb). They are well marked on streamline charts but are usually cloud free. They have not been adequately described in the literature but there are indications that they can interact with tropical cyclones to cause variations in their intensity or rainfall.

3.12 Tornadoes and waterspouts

The man-in-the-street often confuses

Tornadoes and waterspouts ~~are often confused~~ with tropical cyclones, ~~and~~ it is therefore worthwhile elucidating the differences between them. Tornadoes, waterspouts and dust devils are atmospheric vortices or whirlwinds as also are tropical cyclones but the latter are more than one thousand times larger and have much longer lives. The differences between tornadoes, waterspouts and dust devils are mainly ones of intensity and location. Tornadoes are the largest and most intense of the three and occur mostly over land.

3.12.1 Tornadoes

Tornadoes occur most frequently and with the greatest intensity in the United States of America - where they are often called "twisters" - but they also occur in Canada, Europe, India, Australia, Japan, New Zealand and to a much lesser extent in a few other countries situated outside the tropics. The tornado appears as a funnel-shaped cloud with the widest part of the funnel merging into the base of its dark, parent, cumulonimbus thunderstorm cloud, Fig. 3.   . The vortex is made visible by the cloud which forms as a result of cooling caused by expansion in the low pressure core. Most tornadoes in the U.S.A. rotate cyclonically, that is, in a counter clockwise direction when seen from above but, some rotate in the opposite direction (Kessler 1977). Where the rotating spout reaches the ground large amounts of debris are raised within a cloud of dust and there is frequently a clear space between this dust cloud and the higher cloud of water particles forming the funnel. The destructive winds are usually contained within a radius of 200 m.

It was previously thought that the speed of the wind in tornadoes approached that of sound. Dunn and Miller (1964) give estimated speeds of 200 - 220 m/s but both current theory and observations indicate that the highest wind speeds are about the same as those found in the most severe typhoons that is about 100 m/s (Kessler 1970 and 1977, Fujita 1970). These winds have been determined from careful analysis of the movement of debris as shown on cine films of tornadoes and from doppler-radar measurements (see sect 10.8.14). Dergarabedian and Fendell (1970 and 1971) devised a method to estimate the maximum wind from the height of the base of the parent cloud and the shape of the funnel cloud. They determined that wind speeds up to 94 m/s occurred in the tornadoes whose photographs they analysed and, in the case of a well photographed Florida waterspout, they estimated the maximum wind speed to be ~~as high as~~ 80 m/s. Zrinc et al (1977) measured winds of 92 m/s at an altitude of 630 m and 85 m/s at 1.5 km using doppler radar. Vertical velocities up to 60 m/s have been observed and flying debris has been "centrifuged" outwards from the tornado centre at 13 m/s.

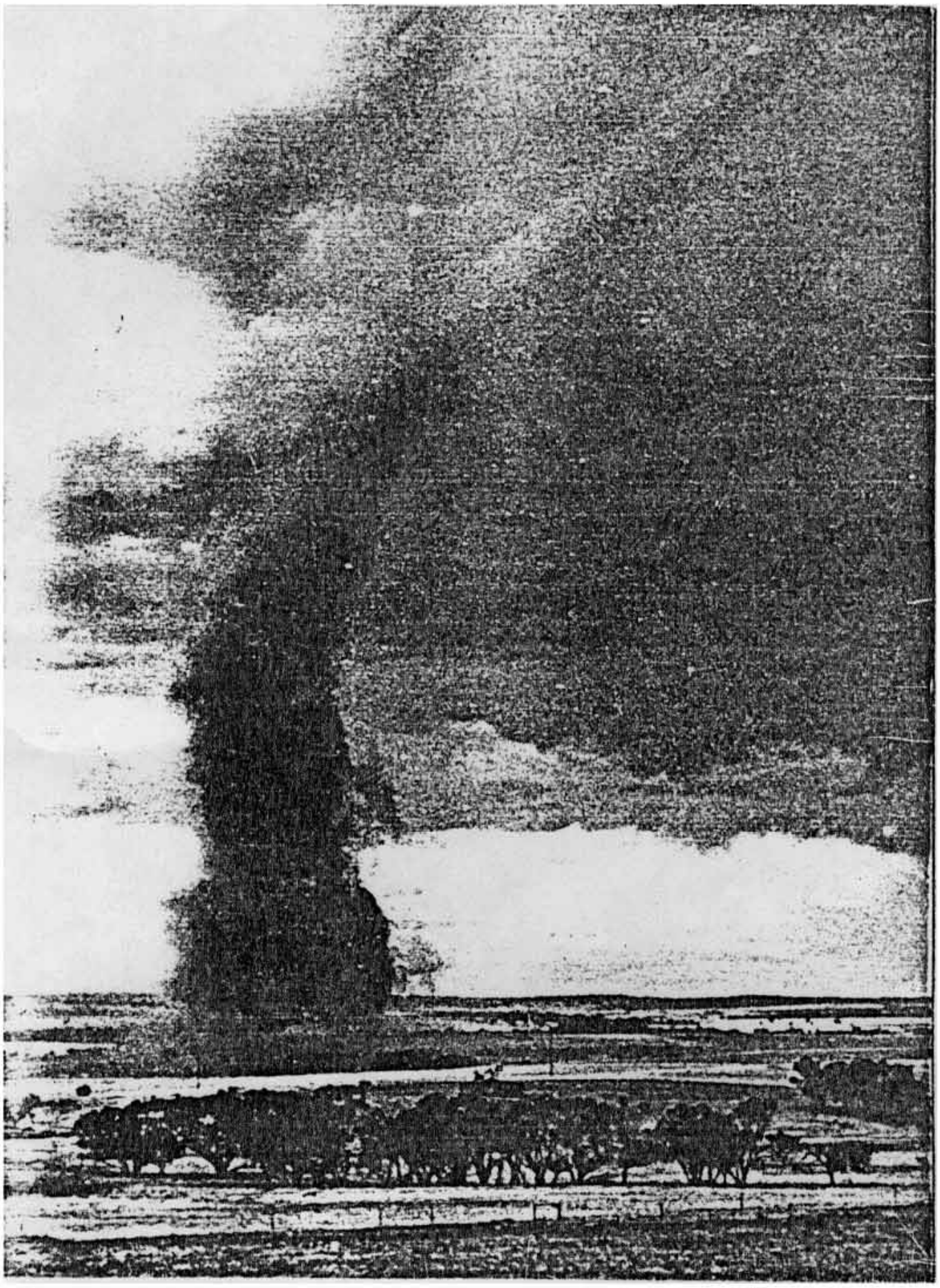
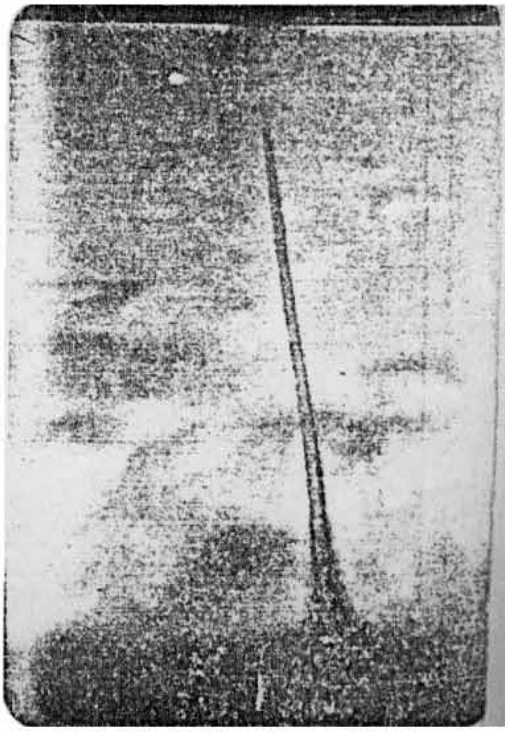


Fig. 3. . A tornado at Northam (80 km northeast of Perth, Australia) on 21 December 1977 (left). The brown dust in the cyclonic (clockwise in the southern hemisphere) vortex column is spiralling hellically upwards towards a small condensation funnel cloud which protruded downwards from the parent cloud. A waterspout in the South China Sea (right) near 14.6°N 118.0°E on 26 April 1969 was also rotating cyclonically. It could not be seen on radar. (Photos(left) by courtesy P.J. May and C.J. Crane and (right) Capt. R.G. McDonald.)



A severe tornado has yet to pass directly over a barograph but an estimate of the central low pressure can be made by noting the height of rise of any water surface beneath the tornado i.e.using the tornado as a water barometer. From these and other considerations it is believed that the lowest pressures generated in tornadoes are probably about the same as in typhoons that is, about 870 mb.

Tornadoes have a life time which is typically between 15 and 60 minutes and cover an average distance of 16 km; the longest known tornado path was 130 km long. They usually move across country at speeds between 0 and 30 m/s reducing many buildings to rubble and lifting light and heavy objects - such as cars - into the air. Extreme damage is caused by tornadoes by virtue of both the force of their winds and the rapid reduction in atmospheric pressure which occurs as the vortex passes. Buildings which suddenly enter the low pressure area near the centre of the core often explode outwards due to the force exerted by the relatively high pressure air inside which is unable to vent sufficiently quickly. The swathe of destruction is, on average, about 400 m wide.

Most tornadoes are axially symmetric. However, when they are large the vortex in the central core region - where suction is greatest - becomes unstable and splits into several vortices which Fujita (1971) calls "suction vortices". Each of these vortices causes its own trail of destruction.

The tornado is sometimes accompanied by a terrific roar the reason for which is not known. One theory attributes the noise to frequent electrical discharges - lightning flashes - within the tornado cloud. Over 70% of tornadoes are associated with electrical activity of some kind and radiate radio waves which have been detected at frequencies between 150 kHz and 200 MHz. These radiations appear to be strongest around 60MHz (Biggs and Waite 1970) at which frequency they can be detected on domestic television channels within a range of about 30 km. The "burst rate" of atmospherics increases with storm intensity. A

Experiments have been made to determine whether the frequency of bursts of radiation received on radios tuned to 3 to 16 MHz can be used to indicate the presence of nearby tornadoes. It is believed that the bursts originate from lightning discharge processes within the parent cloud. Unfortunately, similar bursts of radiation are received from thunderstorms which are not associated with tornadoes and, more seriously, some tornadoes cannot be detected by these means because

A

Visual
1. Observations of lightning from satellites indicate that, on
average, about 3 flashes per minute occur in non-tornado
thunderstorms ~~in~~ the U.S.A ~~and~~ with about 8 flashes per minute
in those with tornadoes.

√

the parent cloud does not generate radio waves. Doppler radar (sect.10.8.4) is a more reliable method of detecting tornadoes and, in addition, indicates both the position of the storm and the speed of its strongest winds.

Electrical theories for the maintenance of the tornadoes have been proposed in the past, but we now know that electrical activity in or near the tunnel is far too infrequent to lend credence to these theories (Davies-Jones and Golden 1975). In any event, we know that winds in excess of 100 m/s have not been observed and winds up to this speed can be explained in terms of horizontal pressure differences created as thermal properties of convectively unstable air are redistributed during overturning (Kessler 1970). However, in the unlikely event of winds significantly in excess of 100 m/s being found in tornadoes then, to explain their origin, some electrical process would need to be invoked.

Tornadoes are always associated with large cumulonimbus clouds which, in turn, are frequently associated with well developed cold fronts and squall lines. They also occur in some regions in association with the inner and outer spiral bands of tropical cyclones (sect. 15.1.). Low-level horizontal wind shear and great instability are necessary for the formation of tornadoes. The instability occurs when a stream of warm, moist air about 1.5 km deep flows under a cool dry atmosphere at intermediate levels brought in by strong winds in the higher atmosphere. These conditions are found, par excellence, in the southern central U.S.A. when southerly winds from the warm waters of the Gulf of Mexico bring warm, moist air to the east of the Rocky Mountains at low levels while, higher up, a jetstream in a wave in the westerlies brings cold air across the mountains and above the warm Gulf air. At such times, systems bringing low level horizontal wind shear - e.g. cold fronts and shear lines - increase the probability of tornado formation. Isaacs et al (1975) have proposed that, in the U.S.A., air dragged along by vehicles on motor ways passing in accord with the right-hand drive rule generates cyclonic shear (Vorticity) which may contribute to the formation of tornadoes. The six-fold increase in the annual incidence of tornadoes in the last 40 years in the U.S.A. and their reduced numbers on Saturdays are cited in support of their theory.

3.12.1 Waterspouts are formed when tornadoes move over a water surface or form there. The low pressure in the funnel cloud causes water to be sucked up to a height of about 1 m - corresponding to the reduction in pressure - and the strong winds pick up spray. A clear space frequently separates the cloud of spray from the funnel cloud (Fig. 3.). Waterspouts, like tornadoes, are usually associated with large thunderstorms and, quite frequently, there will be more than one spout from the parent cloud. In the tropics, weak waterspouts sometimes form over warm seas in relatively calm conditions. I have seen such waterspouts in the autumn months, over the waters around Hong Kong and I have seen stronger ones form in association with large dark cumulonimbus clouds during the early summer. Capt. R.G. McDonald, the master of the m/v "Eastern Maid" (and frequent recipient of the Hong Kong Royal Observatory's observers award) took the photograph in Fig.3. in the South China Sea.

Although some waterspouts are stronger than lesser tornadoes they are usually weaker and smaller than the severe tornadoes of the U.S.A. They rarely last for more than half an hour or travel more than 15 km. They can rotate in either direction.

3.12.2 Dust devils are even weaker than the majority of waterspouts and they can only be identified by the dust and debris they raise or by the spray which they lift as they pass from land across a water surface. They are associated with the convection of hot, dry air and usually form independently of any cloud formation. The strongest dust devils form over hot, dry, uniform deserts in light wind conditions (sect. 3.5.1). Carroll & Byan (1970) found that of 588 dust devils which formed over a test area in the Mojave desert in southern California 314 showed cyclonic and 274 anticyclonic rotation. They concluded that the earth's rotation was not significant in determining the sense of spin of these small vortices which, they considered, was determined by shear between local air streams. They were able to show that shear between rising and descending convection currents was the source of the vorticity for dust devils.

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