



TOM BRADBURY

METEOROLOGY
AND
FLIGHT

A PILOT'S GUIDE TO WEATHER

THIRD EDITION

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Conversion factors

1 m = 3.2808 ft	1 ft = 0.3048 m
1 mm = 0.039370 in	1 in = 25.40 mm
1 km = 0.62137 mile = 0.53993 nautical mile	
1 mile = 1.6093 km = 0.86898 nautical mile	
1 nautical mile = 1.8520 km = 1.1508 mile = 6076.1 ft	
1 metre per sec = 2.2369 mph = 1.9438 nautical mph	
1 ft per sec = 0.68182 mph = 0.59248 nautical mph	
1 mph = 0.86898 knot = 1.6093 km per hour = 0.44704 m per sec	
1 knot = 1.1508 mph = 1.852 km h = 0.51444 m per sec	
1 km per h = 0.62137 mph = 0.53996 knot	
1 kg = 2.2046 lb	1 lb = 0.45359 kg
1 oz = 28.350 grammes = 0.06250 lb	
1 gramme = 0.035274 oz	
1 tonne = 1000 kg = 2204.6 lb	
1 UK ton = 1016 kg = 2240 lb	
1 atmosphere = 1013.25 mbar = 101325 pascal = 760.0 mm Hg = 29.921 in Hg = 33.899 ft water = 14.696 lb sq in = 10332 kg sq m	
1 pascal = 1 newton sq m = 0.01 mbar	
1 millibar = 100 pascal = 0.02953 in Hg = 0.000750 mm Hg	
1 in mercury (Hg) = 33.864 mbar	
1 kW = 1.3410 hp = 1.3596 metric hp (cv)	
1 hp = 745.7 kW = 1.0139 metric hp	

Introduction

This book is chiefly for those who fly for enjoyment and would rather look at the view than concentrate on the instrument panel. It is not intended as a text book on Meteorology, but as a guide to some interesting features of the weather which affect flying.

Everyone needs some background knowledge to interpret what is seen. Some pilots, especially those who fly both with and without an engine, seem to have an amazing ability to interpret features in the sky which many meteorological 'experts' do not see at all. What is observed depends to a large extent on how the observations fit into a mental framework. It is said that when sight was restored to a blind man he was shown an orange but was unable to recognise it until he had put out his hand to feel its roundness and texture. The eye could produce an image but the brain did not have the experience to interpret it.

This problem is not confined to the newly sighted. Some city dwellers do not notice what is happening in the sky above until a shower descends on them. Student pilots occasionally fail to recognise their home town when looking at it from above for the first time. Quite a few meteorologists only know a cloud by its statistical properties, not as a usable source of energy. Some landscape artists draw realistic clouds, while others put in mere splashes of colour to fill in the sky.

In any subject there are dull parts and complicated sections and weather is no exception. What one person finds tedious, another may find interesting. Sections which seem to be over technical can be skipped; the essential points are repeated where necessary in other chapters. This is why some items are to be found in several different chapters.

The appendices contain details of various Met. broadcasts which may be used to keep up with recent weather reports. These range from facsimile and radio teletype broadcasts, which require dedicated equipment, to plain language broadcasts.

Much space has been devoted to features of particular interest to those who fly without an engine. For these pilots even small movements of the air and subtle changes in the clouds are important. Pilots of powered aircraft with modern radio and electronic aids may, when they have to, go from take off to landing without seeing anything in between. To these experts weather is usually of secondary importance. Even so, there are some types of weather which everyone prefers to avoid, and one may need to fly to an airfield without VOR, ILS or radar.

It is hard to make any statement about the weather without adding

qualifying phrases such as ‘...on most occasions...’, ‘...provided that...’, ‘...except when...’. This becomes wearisome to write and exasperating to read. However, the reader needs to be a little cautious in accepting any weather ‘laws’, especially when they are asserted with undue confidence. There are nearly always exceptions. Mathematical theories have achieved remarkable success in producing numerical forecasts but they cannot yet handle all the complexities of the weather. Honest observations are still important.

A note on units

As far as possible the units are those most often used in flying. These are an unscientific mixture agreed by the aeronautical authorities, who were constrained by the fact that many instruments were already calibrated in them when the subject was debated.

Speed is given in knots, height in feet, temperature in degrees Celsius. Distances are mostly given in nautical miles since this is convenient for navigation over a globe divided into latitude and longitude. However, visibilities are reported by Air Traffic Control in metres or kilometres. Pressure is quoted in millibars, but the hectopascal (hPa), which is equivalent to a millibar, appears on many charts issued by European Met. Centres.

Illustrations

The satellite pictures have two sources. The very high resolution pictures (those which have fine detail) were supplied by the University of Dundee Satellite Laboratory. Dundee produces the most magnificent results!

The low resolution pictures (those with lines of latitude and longitude added) were photographed from a video monitor using a WSR 524 bought from Feedback Instruments of Crowborough. This has the great advantage of automatic reception. An internal computer adds the geographic grid to each picture and retains it in memory until the next picture arrives.

The black-and-white photographs have been reproduced from colour slides, mostly taken from the cockpit of sailplanes and powered aircraft.

I

Air in motion

Where the power comes from

The atmosphere is an enormously energetic system. Practically all the power to run it comes from the sun whose radiation provides about 1.4 kW per square metre above the atmosphere. This energy does not all reach the surface of the earth; some of it is immediately reflected back into space. The proportion which is available to warm the surface varies across the globe. Equatorial regions, where the sun passes almost directly overhead, collect most of the energy. Polar regions, where the sun barely rises above the horizon, only receive a significant amount of energy during the summer months, and even then much of this is immediately reflected back into space by the cover of snow and ice.

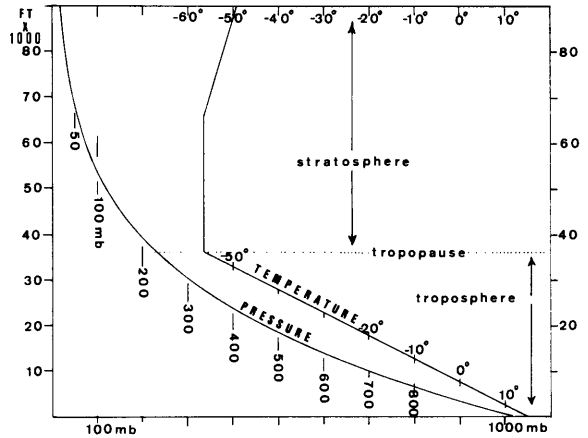
The great contrast of temperature between tropical and polar regions produces an ever-changing flow of wind which redistributes the heat energy and produces the weather systems. The winds are a fundamental part of the weather process and this opening chapter describes something of the wind-producing mechanism.

First, a reminder of some basic features

Most people are familiar with the weather presentation on TV, or see little weather maps in the daily press. The section below may help to remind readers of some important features of the weather.

Pressure

Pressure decreases with height. The rate of change is shown by the lower (curved) line in fig. 1. The most commonly used unit is the millibar (mbar) but recently it has been replaced by the pascal which is only 1/100th of a mbar. Since this is inconveniently small some Met. services now label their charts in hectopascals (hPa) which are exactly the same as millibars.



1. How pressure and temperature change between the surface and 90,000 feet

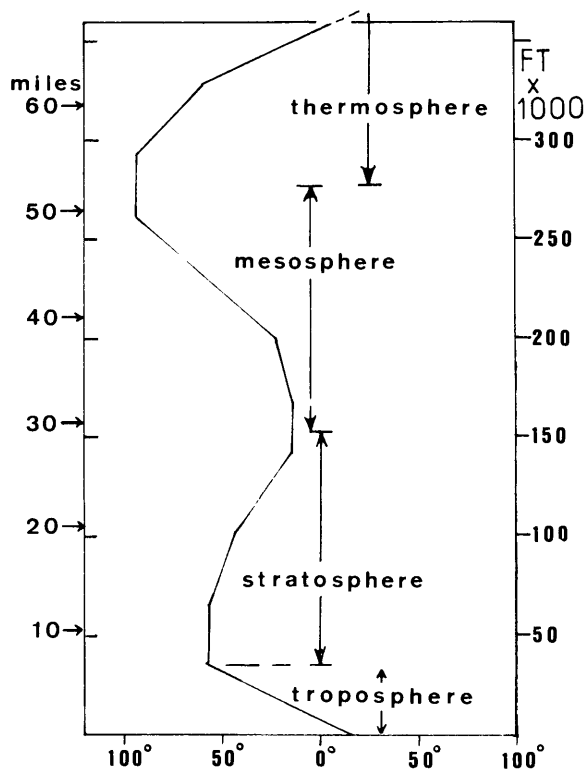
Table 1. Pressure units

- 1 Atmosphere = 1013.25 millibars (mbar)
- = 1013.25 hectopascals (hPa)
- = 29.921 inches of mercury
- = 760.00 mm of mercury
- = 14.696 lbs per square inch
- = 1033.2 kg per square metre

At sea level a pressure change of 1 millibar occurs if you ascend 27 feet. Higher up this figure increases. At 3,000 feet each millibar equals a climb of 30 feet. At the height most jet aircraft fly (around 35,000 feet) you have to ascend about 90 feet to get the same pressure change.

The atmosphere

Figure 1 shows how temperature and pressure vary in the average atmosphere up to 90,000 feet. Figure 2 shows the temperature changes up to a height of 70 miles. These diagrams also show the different layers; each layer has a particular temperature structure.



2. Atmospheric layers and the temperature profile up to 70 miles

The troposphere

Practically all the features which combine to produce 'weather' occur in the lowest layer of the atmosphere called the troposphere, where the average air temperature decreases about 2°C for every 1,000 feet of height. This layer is (on average) about 7 miles deep but it can vary between about 5 and 10 miles and nearly all the clouds and moisture are confined within it. Even the great storms which extend for a 1,000 miles or more are restricted to this relatively shallow layer of the atmosphere.

The stratosphere

Above the troposphere is a layer in which air temperature increases slowly with height. In this region the air is normally far too dry for any clouds to appear.

The tropopause

The boundary between the troposphere and the

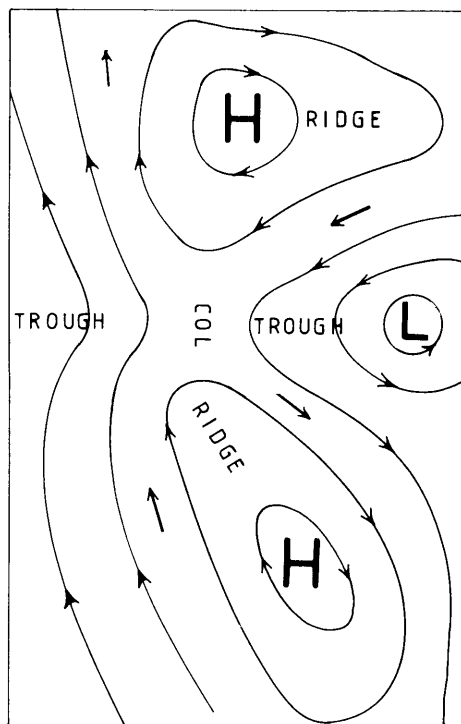
stratosphere is called the tropopause. If you look out from a high-flying aircraft the tropopause can sometimes be seen as a level haze top, or a smooth flat cloud top, at about 36,000 feet. The level is not constant but undulates slowly up and down as major weather systems move beneath it.

The mesosphere and thermosphere

These are the two layers above the stratosphere. The temperature decreases in the mesosphere but increases with height in the thermosphere. At higher levels in the thermosphere the temperature becomes extremely high. However, there is so little air there that these high temperatures have no practical effect on the few astronauts who pass through them.

Isobars

These are lines of equal pressure. They are drawn on many weather maps to show up the different pressure systems such as anticyclones (highs), depressions (lows) and the associated valleys (called troughs) or their opposites (the ridges). Figure 3 illustrates these features.



3. Isobaric patterns

Isobars and the wind

Over most of the globe, with the exception of regions near the equator, the wind blows almost parallel to the isobars. The strength of the wind depends on how close the isobars are spaced. Close spacing indicates a strong wind, wide spacing produces a light wind.

Buys Ballot's law

This states that in the Northern Hemisphere, if you stand with your back to the wind the lowest pressure lies on the left hand side. In the Southern Hemisphere, conditions are reversed and low pressure lies to the right when your back is to the wind.

Streamlines

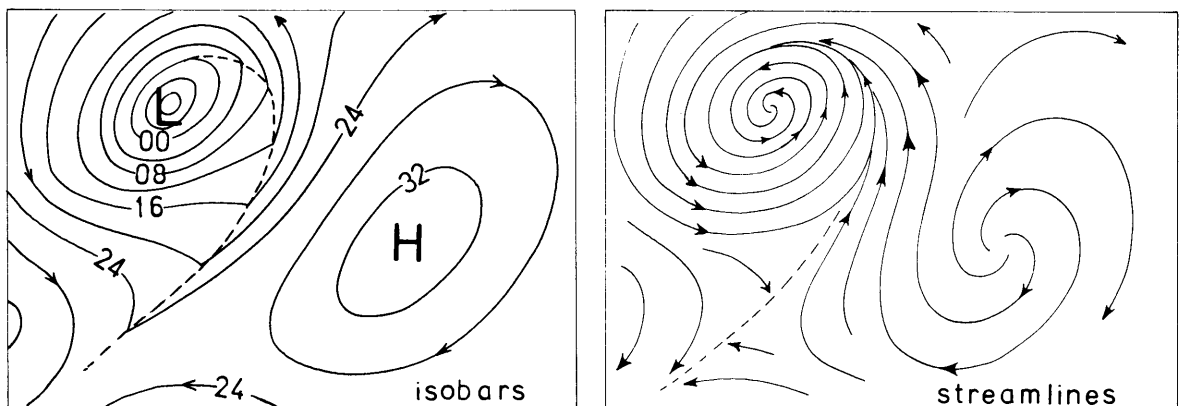
In tropical regions, and particularly near the equator, the isobars no longer give an accurate indication of the wind. They are replaced by streamlines, which follow the direction of flow of the wind. Streamlines are not so easy to draw as isobars but they can show where the air flow is coming together (converging) or spreading apart (diverging). A comparison between an isobar chart and a streamline chart is shown in fig. 4. You may notice that the streamlines spiral out from the high and in towards the low and that several streamlines converge on the dotted line marking a frontal trough. By their nature isobars cannot meet in this way.

1940s when the availability of a large number of aircraft and balloon soundings of temperature made it possible to map the flow at high levels. To draw isobars one uses observations of pressure which have been corrected to give the value at some constant level, usually sea level. To draw contours one selects a convenient pressure and draws contours to show how the height varies. For example, the 1,000 mbar pressure surface is usually close to sea level. The actual height depends on the sea-level pressure and the air temperature. If one assumes a temperature of 15°C the relationship between contours and isobars is shown in Table 2 (below).

Table 2. Isobars and the 1,000 mbar contour height

mbar	metres	feet
960	-344.5	-1,130
970	-253.7	-843
980	-174.0	-571
990	-84.8	-278
1,000	0	0
1,010	84.0	275
1,020	167.1	548
1,030	249.5	818
1,040	331.0	1,086

The difference in height shows how the surface slopes. Table 2 also shows the kind of altimeter error one may expect when flying across the wind from high to low pressure. In the extreme case of a flight



4. Comparison between isobars and the streamline pattern

Contour charts

The pattern of winds can also be represented by contour charts. These came into general use in the

from an anticyclone of 1,040 mbar to a deep low of 960 mbar the true height would be more than 2,200 feet lower than that indicated.

Using contours instead of isobars

The height of a pressure surface depends both on the sea-level pressure and the temperature of the air. The air temperature becomes a major factor at high levels. This can be observed when flying over the sea by comparing the height indicated by a pressure altimeter with the true height measured by a radio altimeter. For example, an aircraft flying at an indicated altitude of 30,000 feet (where the pressure is almost 300 mbar) would have a true altitude of about 32,000 feet in tropical regions (where the air below is warm). Flying towards the poles the radio altimeter would show a gradual decrease in height as the air below became colder. Eventually, the true altitude would be about 28,000 feet.

The change in height is due to the slope of the 300 mbar surface which the aircraft was following. Air expands when heated, so that a pressure surface rises when the air is warmed and sinks when the air is cooled.

Contours and the wind

Contour charts can be used just like the more familiar charts of isobars. The winds aloft blow almost parallel to the contour lines. Buys Ballot's law still holds. If you face down wind the lower contour heights lie to the left (in the northern hemisphere). The strength of the wind depends on the steepness of the slope. Close spacing of contours indicates a steep slope and consequently a strong wind. A wide spacing

between contours shows a gentle slope which only produces a light wind.

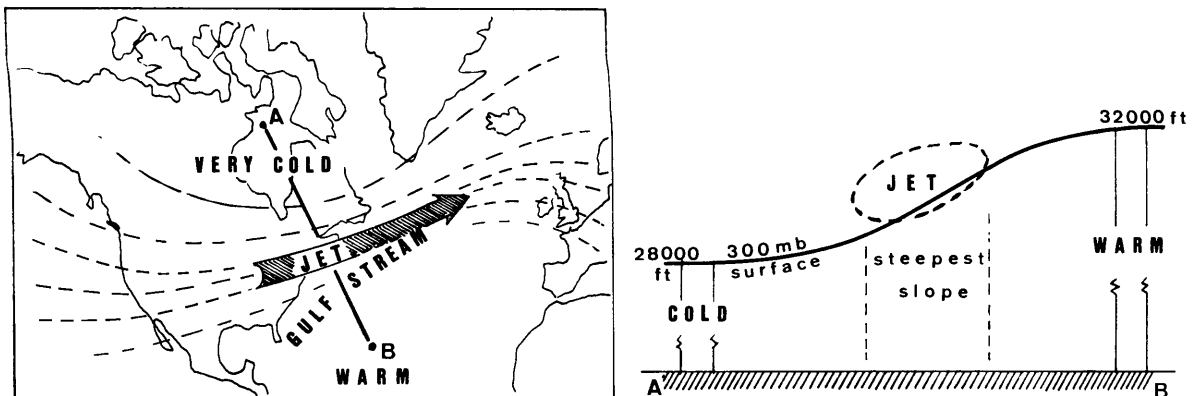
Jet streams

Jet streams are bands of very strong winds which are usually found at levels between 30,000 and 40,000 feet, but which may appear at other heights. A jet stream is often more than a thousand miles long, one or two hundred miles wide and generally less than four miles deep. Most jet streams occur where there is a marked contrast in temperature, for example, where cold air from polar regions is brought close to much warmer air from the tropics.

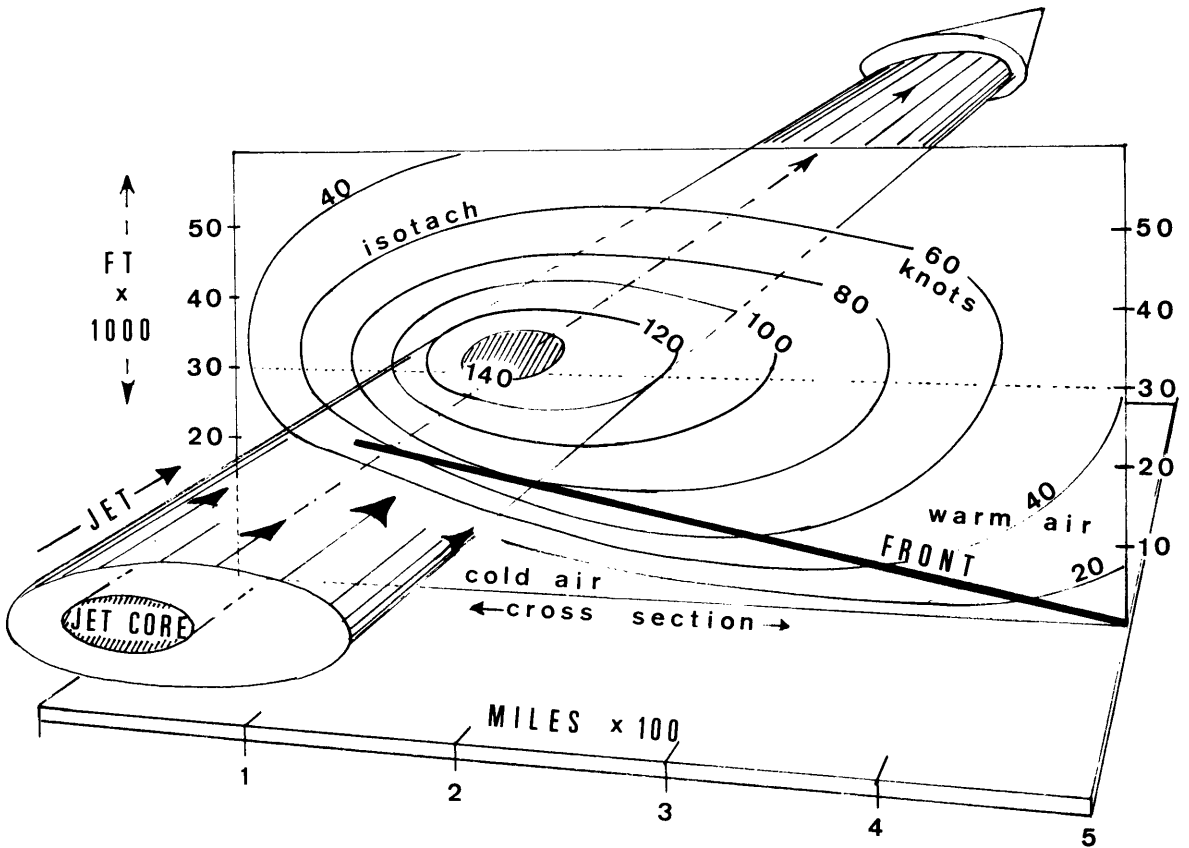
Figure 5 shows an area close to America where jet streams often form, particularly in winter. The strong contrast in temperature makes the contours come very close together, indicating a steepening of the slope, the wind speed increases.

The lower part of fig. 5 shows a cross section along the line A—B. The 300 mbar surface dips down towards the cold air and the jet stream occurs where the slope is greatest. Speeds of 100 knots are common, and speeds of more than 200 knots have been observed in the central core of some powerful jets. Once a jet has developed it can move off down wind and travel thousands of miles, snaking up towards polar regions and dipping down towards the edge of the tropics.

Figure 6 shows a 3-D diagram of a jet stream. The series of rings are isotachs which show how the wind speed increases towards the core of the jet. A jet is



5. Area where jet streams develop and the cross-section of the 300 mbar surface showing the steepening at a jet stream



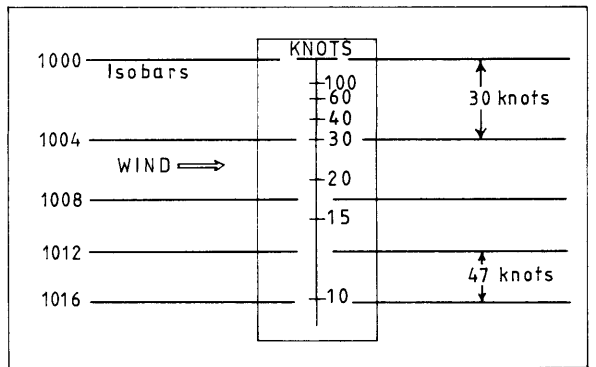
6. Cross-section of a jet stream

often associated with a 'front' (shown on this diagram by the sloping black line). Jets play an important part in the development of depressions. This is described in more detail in Chapter 2.

Geostrophic winds and how to measure them on a chart

The wind speed measured from a chart of isobars or contours is known as the geostrophic wind. It is usually a good approximation to the real wind but there are occasions (mentioned further on) when the actual wind differs significantly from the geostrophic wind. Meteorologists use a transparent scale (called a geostrophic scale) which is laid at right angles to the isobars. The wind speed is found from the spacing between isobars (see fig. 7).

Geostrophic scales are often printed near the corner of a weather map; they are more complicated than fig. 7 because they have to take account of variations



USING A GEOSTROPHIC SCALE FOR 4 MB ISOBARS

7. Example of a simple geostrophic wind scale and its use

of map scale and latitude. Each size of chart needs a different scale which is constructed for a particular interval between isobars. For example, in England isobars are usually drawn at intervals of 4 mbar. However in France and Germany one commonly sees

a 5 mbar interval. This makes life difficult for anyone who travels to different countries. The problem can be avoided as follows.

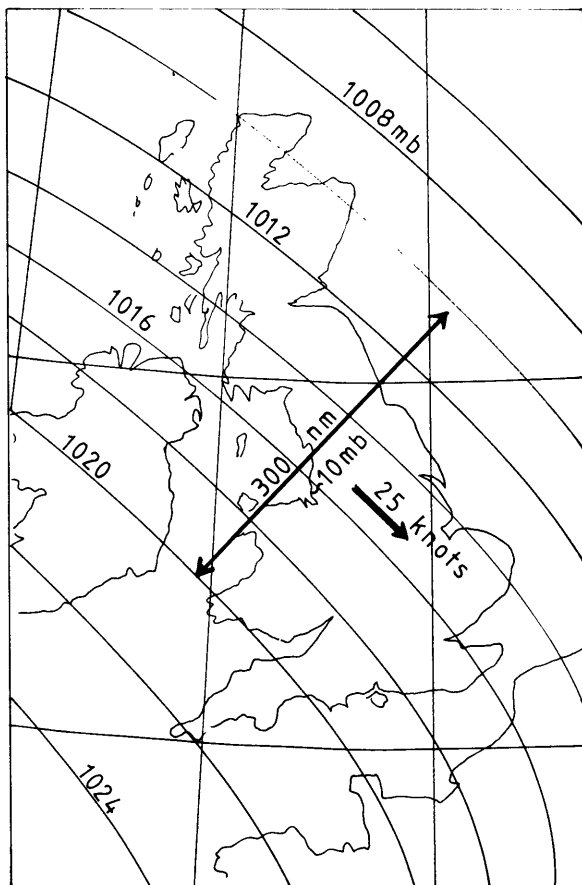
Finding geostrophic winds without a scale

The process is illustrated in fig. 8.

(a) From the weather map measure off a distance of 5 degrees of latitude. This is equivalent to 300 nautical miles. Nearly all maps have lines of latitude marked on them. One can use a pair of dividers, mark a transparent piece of plastic or even mark the edge of a piece of paper.

(b) Lay the marked-out length at right angles to the isobars or contours in the area of interest and note the change in pressure or height between the ends.

(c) Multiply this by the factors in the table below to get the wind speed in knots.



8. Illustration of how to find a geostrophic wind without a scale

Table 3. Factors to derive the geostrophic wind speed in knots from the change in pressure or height over a distance of 5 degrees latitude (300 nm)

Latitude degrees	Multiplication factors	
	For isobars (mbar)	For contours (metres)
70	2.1	0.25
60	2.3	0.27
55	2.4	0.29
50	2.6	0.31
45	2.8	0.33
40	3.1	0.37
35	3.4	0.41
30	3.9	0.47
25	4.7	0.56

For example, if the surface chart (drawn in millibars) had a 10 mbar drop at latitude 52 degrees where the factor in the second column is about 2.5, then the wind speed would be $10 \times 2.5 = 25$ knots.

If instead of isobars the chart showed contours one would use the factors in the third column. Thus, if the contours showed a height change of 200 metres and the latitude was 45 degrees then (using the factor 0.33) the wind speed would be $200 \times 0.33 = 66$ knots.

Notice that as the latitude becomes less the multiplication factor becomes larger. Thus, a contour spacing which would give a wind speed of 50 knots at latitude 70 would give a wind speed of 94 knots at latitude 25. In low latitudes (towards the equator) where the factor becomes very large, the wind is seldom truly geostrophic and this kind of measurement becomes inaccurate.

How the wind changes near the ground

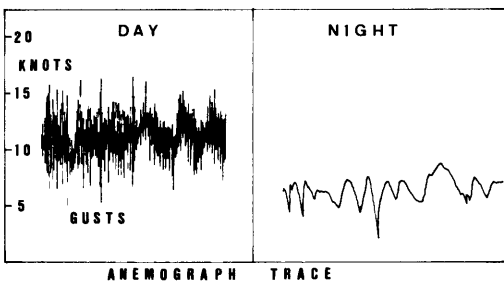
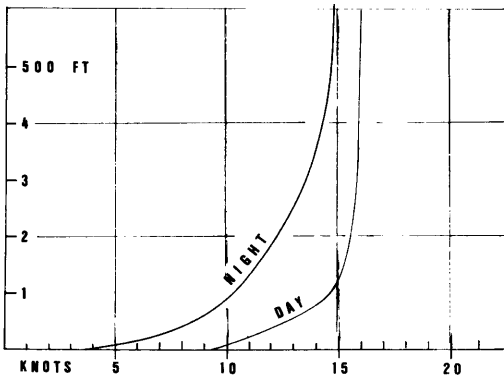
Geostrophic winds, whether they are measured from isobars or a contour chart, do not allow for any friction. The wind near the ground is slowed down by friction; the rougher the surface the greater the friction. It is generally assumed that the wind measured from the surface isobars will be about right at levels of 2,000 feet. Below that, the effect of surface friction slows the wind down. The wind direction then turns to blow across the isobars towards lower pressure.

There is no single relationship between the geostrophic wind and the surface wind; it depends on the surface and the time of day. For example, if the 2,000

foot wind was 270 degrees 20 knots then the surface wind would be about:

- 260 degrees 15 knots over a smooth sea
- 250 degrees 9 knots over land by day
- 240 degrees 7 knots over land at night with some clouds
- 230 degrees 5 knots over land on a cloudless night

Figure 9 illustrates how the wind speed may vary with height in the lower layers. Notice that the speed changes most rapidly close to the ground. The change of wind speed is very noticeable if you go across a high suspension bridge. When gales are blowing, high-sided vehicles such as furniture vans may have little trouble travelling along country roads with trees and hedges to reduce the wind speed; however, on a high suspension bridge the full strength of the gale may blow the vehicle over on to its side. Aircraft landing in strong winds need extra airspeed to compensate for the sudden drop in wind speed close to the ground. The lower part shows anemograph (wind recorder) traces for day and night. By day the trace shows hundreds of gusts and lulls over a few hours. At night the speed is lower and changes much less rapidly.



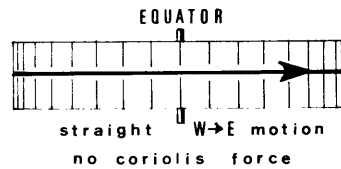
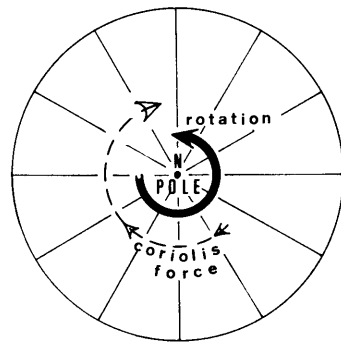
How the geostrophic wind develops and why it can differ from the real wind

Many people do not need to bother with the details of the geostrophic wind. Unless you are particularly interested in the subject the rest of this chapter can be skipped.

Introducing the coriolis force

If the earth did not rotate on its axis there would be no geostrophic wind. The air would move directly towards a region of low pressure or flow down any slope shown by the contours. This would tend to prevent the deepening of nearly all depressions. However, the rotation of the earth deflects the moving air. This deflecting force is called the 'coriolis force'.

The upper part of fig. 10 represents the view looking down at the earth from directly above the north pole. The earth rotates in an anticlockwise direction as shown by the thick black arrow. When the air moves, the earth rotates under it. To an observer in space the air appears to be moving in a straight line but to an observer on the ground the air follows a path which curves to the right (shown by the pecked line).



10. Rotation of the earth and the coriolis force

An observer high above the equator sees the earth's surface moving straight from west to east with no apparent rotation. No rotation means no coriolis force and no geostrophic wind on or near the equator.

9. The variation of wind speed with height and what anemograph traces look like by day and night

To an observer above the south pole the earth's rotation is clockwise so the coriolis force produces a deflection towards the left in the southern hemisphere. The effect of the coriolis force is greatest near the poles; it varies as the SINE of the latitude to become zero at the equator.

How geostrophic balance is reached

The geostrophic wind is achieved when two forces are in balance; these are:

(a) the force of gravity pulling the air down a sloping pressure surface (or the pressure force when using isobars) and

(b) the coriolis force tending to turn the air to the right (in the northern hemisphere).

Figure 11 illustrates how the balance is reached. If a previously level pressure surface was suddenly tilted, the air would start to flow down the slope under the pull of gravity. As soon as it started to move the coriolis force would begin to deflect it. It would take several hours for this deflection to have its full effect. The effect is more rapid near the poles than at low latitudes. Eventually the path of the air is turned through a right angle so that it flows parallel to the slope (or to the isobars). The coriolis force still tends to deflect the wind but if the path of the air were turned any further it would start to flow uphill. Instead, a balance is reached when the pull of gravity is exactly balanced by the coriolis force. Now we have a geostrophic wind.

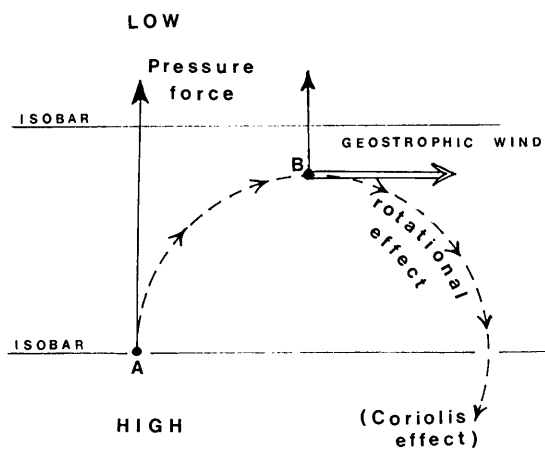
To summarise: the geostrophic wind blows parallel to isobars or contours with low pressure (or height) to the left in the northern hemisphere. The wind speed depends both on the gradient and the latitude. When the gradient is steep the wind is strong; for a given gradient the speed increases as the latitude decreases. The relationship breaks down near the equator where the coriolis force vanishes.

When the wind is not geostrophic

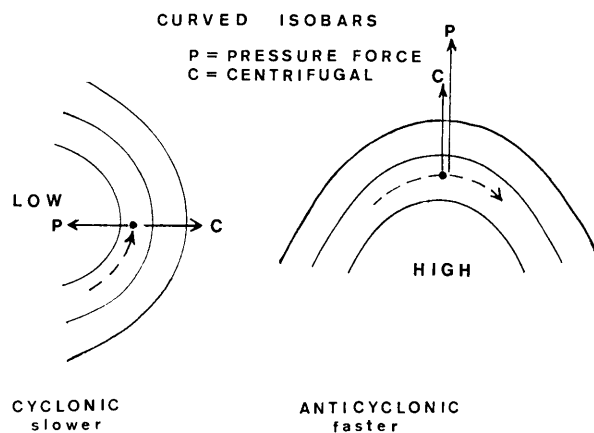
If the wind was always perfectly geostrophic most of the familiar weather systems would never develop. In fact, the geostrophic balance is easily upset and when this happens the air currents can converge and produce areas of ascent where cloud and rain develop.

The gradient wind

The gradient wind is a modification of the geostrophic wind to take account of curved flow. When the isobars/contours are curved a third force is introduced; this is the centrifugal force, familiar to all who drive round corners rapidly. The effect on the wind speed depends on whether the curve is cyclonic (round a low or trough) or anticyclonic (round a high or ridge). Figure 12 shows the two possibilities in the northern hemisphere where cyclonic curves are those which turn to the left.



11. *The development of geostrophic balance*



12. *Illustrating the effect of curved isobars on wind speed*

With cyclonic curvature the pressure force (P) acts inwards but the centrifugal force (C) acts outwards. As a result the effect of the pressure force is reduced and so the wind speed is reduced. With a very small radius of curvature (about 100 miles) the wind may be reduced to half the speed for straight isobars.

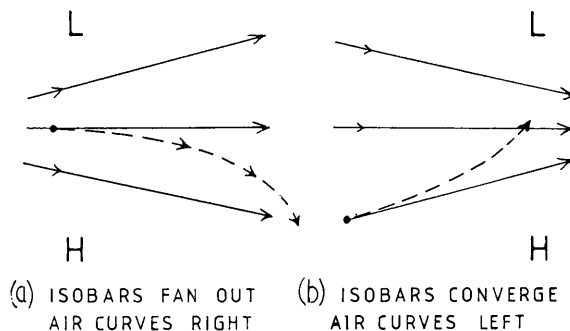
When the curvature is anticyclonic both the pressure force and the centrifugal force act in the same direction (outwards). This increases the effect of the pressure force and so the wind becomes much stronger. The wind flowing round a marked ridge may be increased to nearly twice the speed for straight isobars. This is the upper limit, beyond it the flow ceases to be steady.

Looked at another way, one cannot have closely-spaced isobars near the centre of a high because the wind speed would exceed the limit for an anticyclonic curve and the air would slide outwards instead of following the bend. This would bring down the pressure in the high and reduce the gradient to permitted limits. It is different round a low; if the air slides outwards it will reduce the pressure in the centre and increase the gradient.

When the lines are not parallel

If the isobars/contours are not parallel the geostrophic balance is upset for a different reason which may be summarised as momentum (or the lack of it, inertia). At sea level the mass of air above every square mile weighs about 26 million tons. This mass takes time to respond to a change in the pressure gradient or an alteration in the slope of contours.

Figure 13 shows two possibilities. In (a) the lines fan out; the gradient is decreasing down wind. The air enters from the left with considerable momentum



13. The effect of isobars which fan out or converge

and the speed is soon too high for the gradient. The pull to the left becomes less as the lines fan out but the coriolis pull to the right remains the same until the air slows down. For a time the coriolis force is the stronger and the air turns across the isobars towards the right. The faster the air is travelling when it reaches this region the more noticeable is this right-hand deflection. It quite often amounts to 20 or 30 degrees at the levels where long distance aircraft fly.

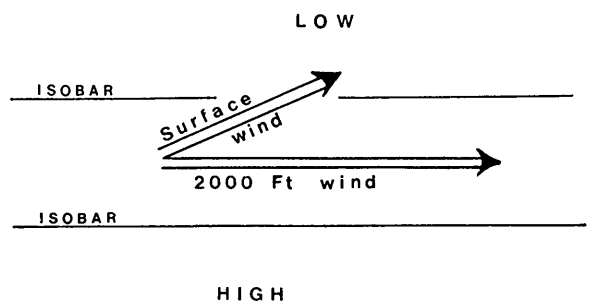
The opposite effect occurs when the lines converge down wind. Then the inertia prevents the air accelerating fast enough. Since the air is moving too slowly the coriolis force is too weak. The strengthening pressure force then takes the air across the isobars towards the low, as shown by the pecked line.

Rapid pressure changes

A similar effect can be noticed when there is a rapid fall of pressure. Once again the wind fails to respond to the new conditions and the air starts to flow across the isobars towards the region where the pressure is falling most. The component across the isobars is called the 'isallobaric wind'. Its effect is most marked when the original wind speed is fairly light. Then the rapid fall of pressure (ahead of an advancing low or trough) can make the wind direction swing 45 degrees or more and turn inwards towards the trough.

Friction

Frictional drag slows down the air near the ground. This upsets the geostrophic balance. The slower the air moves the less effective is the coriolis force. The result is that the slower air is deflected towards the low pressure side, shown in fig. 14.



14. The deflection of surface winds due to frictional drag

Convergence and divergence

Winds which differ from the geostrophic are termed 'ageostrophic'. When they occur the flow of air develops regions of 'convergence' where the streamlines (but not the isobars) come together, or 'divergence' where the streamlines move apart. Convergence is rather like a line of traffic slowing down as it approaches a road junction or an obstruction. After a time all the vehicles are nose-to-tail and the load on the road is increased. Divergence is the opposite effect; once past the obstruction cars speed up and the traffic becomes well spaced out.

If the air started to converge at all levels the excess would cause a rise of pressure at the surface. When

traffic approaches a hold up some drivers may turn off down a side street rather than join the queue. In the atmosphere convergence at one level often results in some of the air moving to another level where the air is diverging.

When air flow converges at low levels but diverges higher up some of the air is forced to ascend into the divergent region. Air cools as it rises and cooling eventually leads to condensation of moisture to form clouds and rain. This can be the start of a major bad weather system. The process, which is often initiated by an acceleration in the fast-moving flow at levels some 4 to 8 miles up, is a major factor in the formation of large depressions and anticyclones. The process is described in Chapter Two.

The development of depressions and anticyclones

Depressions are responsible for much of the bad weather which occurs; this section is an account of some of the reasons why areas of low and high pressure develop. Two important factors are:

(a) a large supply of heat to provide the energy to transport huge masses of air, and

(b) regions where the wind is no longer geostrophic so that the air converges in one place and diverges at another.

Making a little depression

This process can start from calm conditions both at the surface and up aloft. It works best over land when there is strong sunshine. Figure 15 illustrates this.

(a) Sunshine warms the ground and heat is then transferred to the air above. If conditions are suitable this heat is carried upwards by convective currents so that a broad column of air (say 50 miles in diameter) becomes warmer than its surroundings.

(b) This warm column expands and pushes up the pressure surfaces to form a dome in the contours aloft.

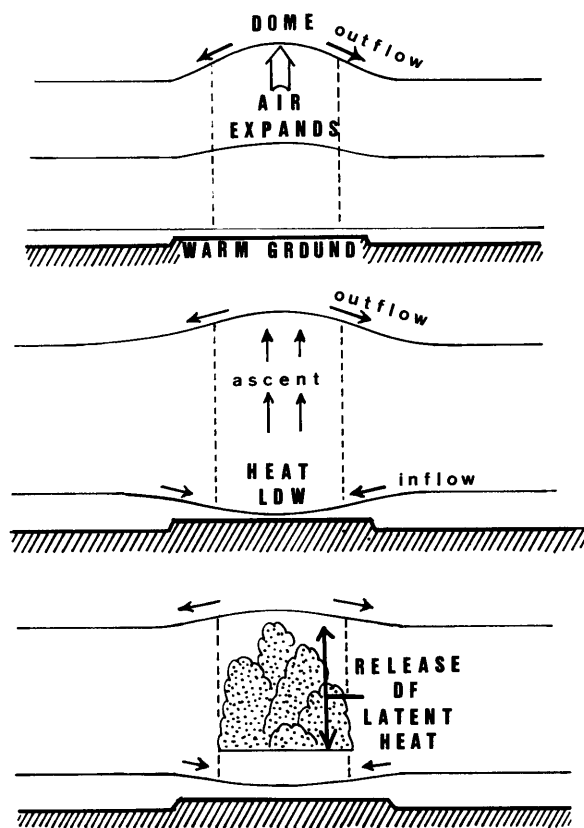
(c) The air forming the dome then starts to move directly downhill under the pull of gravity. Given a quarter of a day the coriolis force would deflect the flow until it became parallel to the slope, but to begin with the airflow is not geostrophic and diverges in all directions.

(d) This divergence aloft reduces the total amount of air below, so the surface pressure starts to fall. The little low has started to form. This is how 'heat lows' form over land on sunny days. It is not the most efficient way of forming a new low because the energy supply is cut off every night but it is effective in tropical regions.

(e) Once there is a surface low the air all round begins to move in to fill it up. Again, the coriolis force takes time to deflect this inward flow and low-level convergence acts to limit the pressure fall.

The process can be improved when the air is moist enough for the rising air to condense and form clouds. Condensation provides an additional supply of energy by releasing latent heat. A wide column of convection cloud can become a sort of chimney. Once the low has formed, inflow at the bottom provides a steady supply of fresh moist air. Ascent releases latent heat which warms the air rising up the 'chimney'. The air eventually spreads out sideways from the top. This is one example of the kind of 'feedback' which often enables an initially weak process to become much more effective with time.

Some 'polar lows' are thought to develop in this



15. Formation of a heat low shown in cross-section

way when very cold air moves out over the relatively warm ocean.

Making a large depression

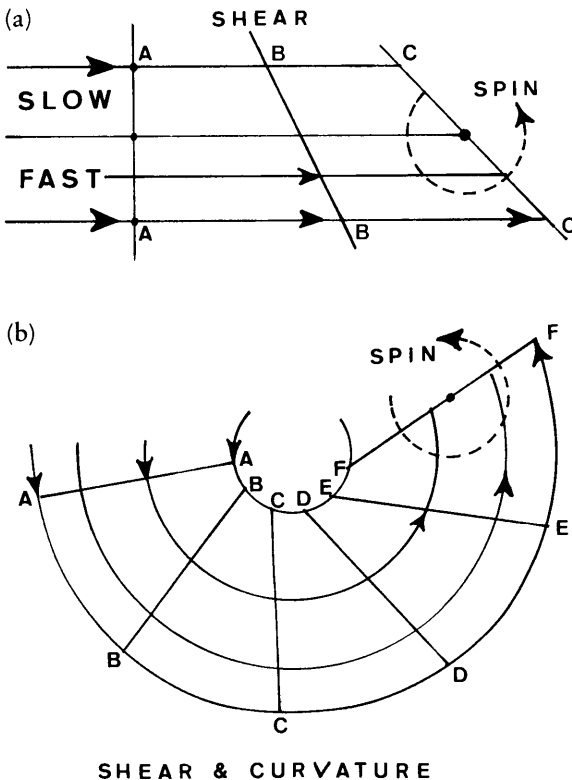
The process described above is not powerful enough to generate a really big depression though some of the features assist in the formation of tropical depressions.

The depression-making process is much more efficient if it is started off by a strongly divergent flow at high levels.

Chapter One ended with a paragraph on convergence and divergence. This kind of flow is closely connected with another feature—'vorticity'.

Introducing the idea of vorticity

Vorticity is the term used to describe the amount of spin. It is made up of two parts, shear and rotation, illustrated in fig. 16. In (a) there is straight flow with



16. Vorticity due to (a) shear and (b) shear and rotation

slower winds on the left of the flow and stronger winds on the right. Suppose an aircraft left a very long-lasting condensation trail along the line A-A. As the trail moved along with the wind it would slowly rotate, first to B-B and later to C-C. This reveals the amount of spin (vorticity) produced just by the shear of wind.

The spin is greater if the air also flows round a curve as in (b). In this case the starting line A-A is eventually twisted round to lie along F-F.

Vorticity and divergence

It is an everyday experience to see the vortex which forms when a basin of water empties down the plug-hole. There is always some slight vorticity in baths or basins and the asymmetric flow towards the plug introduces more. When some water starts to drain out the rest begins to converge upon the exit. The spin, which was originally too slight to be noticed, is increased where the water converges. Soon it becomes so great that a visible vortex forms where the water spins round very rapidly before disappearing down the plug-hole.

The effect is reversed if the water empties into a large container where it can spread out (diverge). The spin then rapidly disappears.

In the atmosphere the most striking examples of vorticity are tornadoes and waterspouts. These are found where the air converges towards the base of a thunder cloud into which the air is being drawn very rapidly. Little dust devils are produced in the same way on hot sunny days when there is a vigorous ascent of air. The difference is that dust devils are much smaller and can develop without a cloud above them.

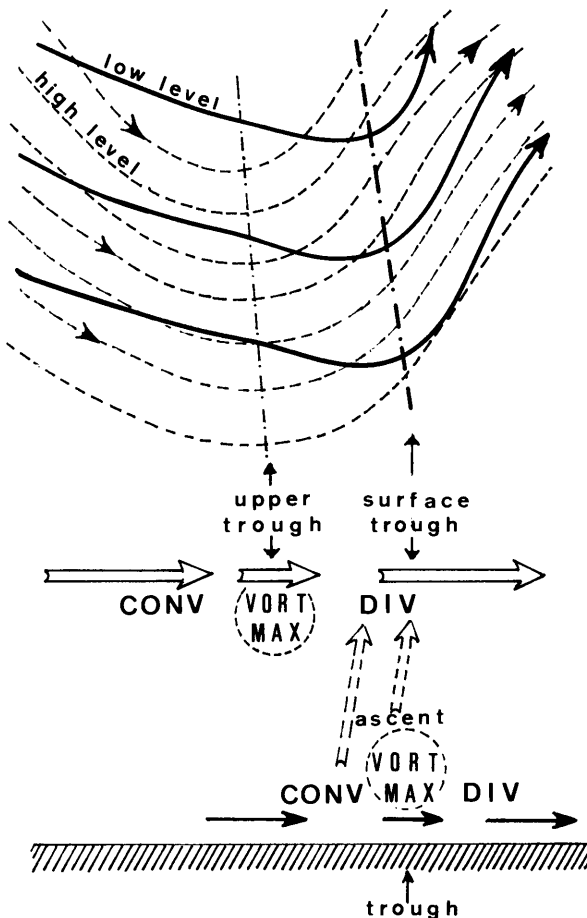
Large depressions which extend for hundreds of miles are sometimes called vortices. They may not seem to have much spin when seen on a single satellite picture but a time-lapse series often shows the cloud swirling round the centre.

Vorticity can be calculated but it is a tedious exercise best left to computers. All that most of us need to know is that vorticity is increased when air flows through a trough and decreased when it flows round a ridge. This implies that air will (nearly always) be converging as it goes through the trough and diverging when it rounds a ridge.

Ascent of air where upper and lower troughs are out of phase

When charts for the flow aloft are superimposed on surface charts one frequently finds that the surface trough lies down wind of the upper trough. This is shown in fig. 17. The surface flow is shown by full lines and the upper flow by pecked lines. The upper trough lies to the left (upwind) of the surface trough. The vorticity is at a maximum along the trough lines (shown on the lower part of the diagram as 'VORT MAX').

As the air approaches this region there is convergence; when it moves on into decreasing vorticity there is divergence. When the upper and lower patterns are superimposed we can see that the convergence at low level lies under the divergence at high level. In this region air is forced to ascend from the low level trough.



17. Ascent of air where surface trough precedes an upper trough. Plan view and cross-section of ascent

Jet streams and the formation of lows

To get strong divergence it is usually necessary to have a fast-moving current of air aloft. In jet streams, where there is a band of air travelling at speeds of 100 knots or more, there is strong wind shear. When this is combined with a curved flow the air experiences big changes of vorticity; it speeds up and diverges in one region and slows down and converges in another. The stronger the winds aloft, the more marked can be this divergence.

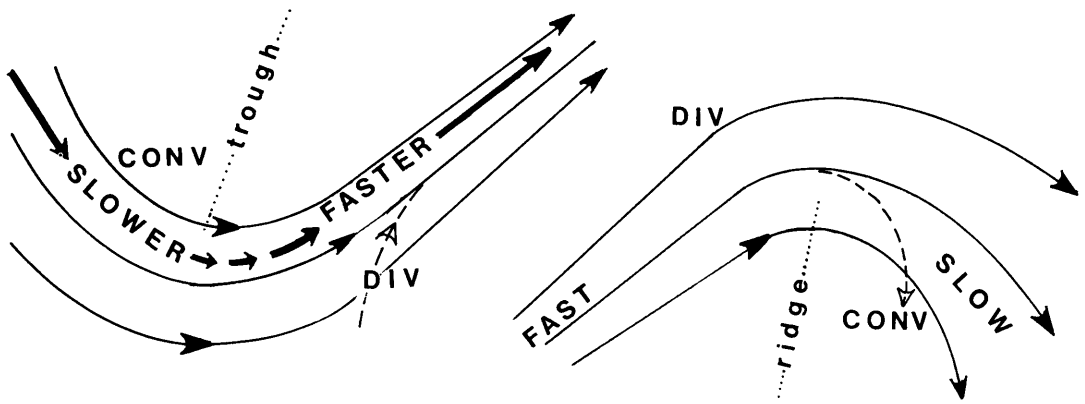
Figure 18 shows a straight section where the contours are close together in the middle indicating a jet stream. As the air approaches the jet it starts to move faster. There is divergence and an inflow along the dotted line from the region marked 'IN' near the jet entrance. At the other end (the jet exit) the air starts to slow down in the region of convergence; the pecked line shows that some air moves across the contours to the region marked 'OUT'.



18. Divergence and convergence in a straight jet stream

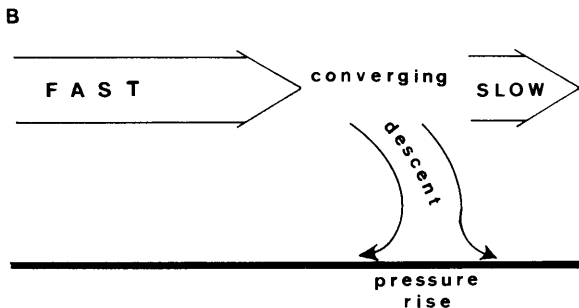
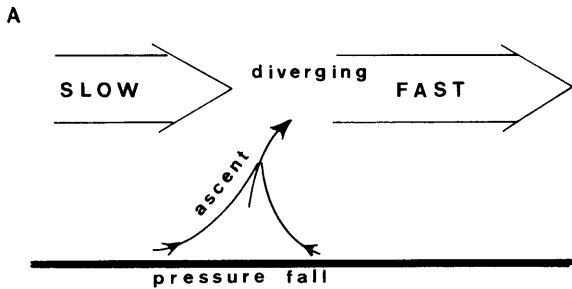
This type of flow has been examined by tracking constant pressure balloons through a jet. The balloons followed tracks similar to the pecked lines. At the jet exit some of the balloons curved so far to the right that they ended up going round in right-handed circles. It is believed that one of the early, unsuccessful attempts to cross the Atlantic by balloon failed because the balloon left the jet exit and got stuck in the right-handed swirl.

Figure 19 shows a jet stream with a trough upstream on the left and a ridge downstream on the right. The curvature adds to the vorticity and increases the divergence near the right entrance of the jet. With a pattern like this it is common to find that a surface depression forms under the region marked 'DIV' while an anticyclone develops in the region below the 'CONV'.



19. *Divergence and convergence when jet stream lies between trough and ridge*

Figure 20 is a vertical cross section illustrating this process. In (a) the ascent of air into the region of upper divergence produces a fall of pressure and low-level convergence near the surface. In (b) the slowing down of the jet in the region of upper convergence results in a rise of pressure at the surface below and descent of air.



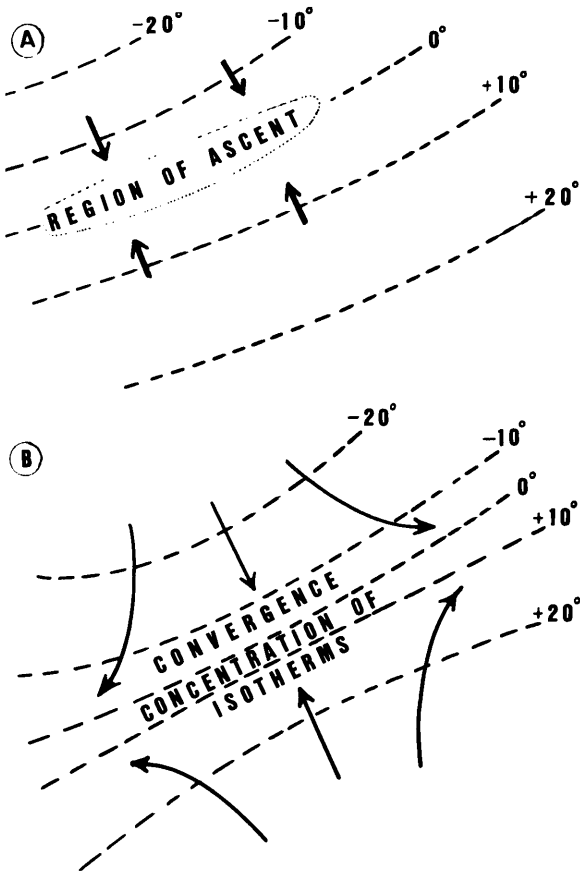
20. *Vertical motion and surface pressure change beneath regions of high-level divergence and convergence*

Effect of low-level convergence on the development of a front

When pressure begins to fall at the surface the low-level air starts to converge towards that region from all sides. Colder air from polar regions is brought closer to warmer air from the sub-tropics. This increases the contrast between the cold and warm air masses. The convergence is maintained by ascent of air from low to high levels; it eventually draws warm and cold air masses so close together that a well-defined boundary forms between them. This boundary is termed a 'front'. The process is illustrated in fig. 21 (a) and (b). Figure 22 shows the cross section of a front. The warm air, being less dense than the cold, lies above it and the frontal surface forms a sloping boundary at an angle of roughly 1 : 100. (The slope is not constant.)

Feedback from low-level convergence

The low-level convergence not only intensifies the temperature contrast between warm and cold air masses, it also steepens the slope of the pressure surfaces aloft and consequently strengthens the jet stream. Here is an example of positive feedback: the jet started low-level convergence, the low-level convergence increased the temperature contrast, and this in turn strengthened the jet.



Adding more energy to the system

When the low-level air starts to ascend, a new source of energy becomes available. The rising air cools and the water vapour starts to condense to form clouds. Condensation releases latent heat. If the rising air is warm it can contain a large amount of water vapour. Each kilogramme of tropical air may hold some 20 grammes of water vapour at sea level. By the time this air has risen 3 miles the cooling will have made half of the vapour condense as clouds, releasing nearly 600 calories per gramme. This adds a lot of extra energy to the system.

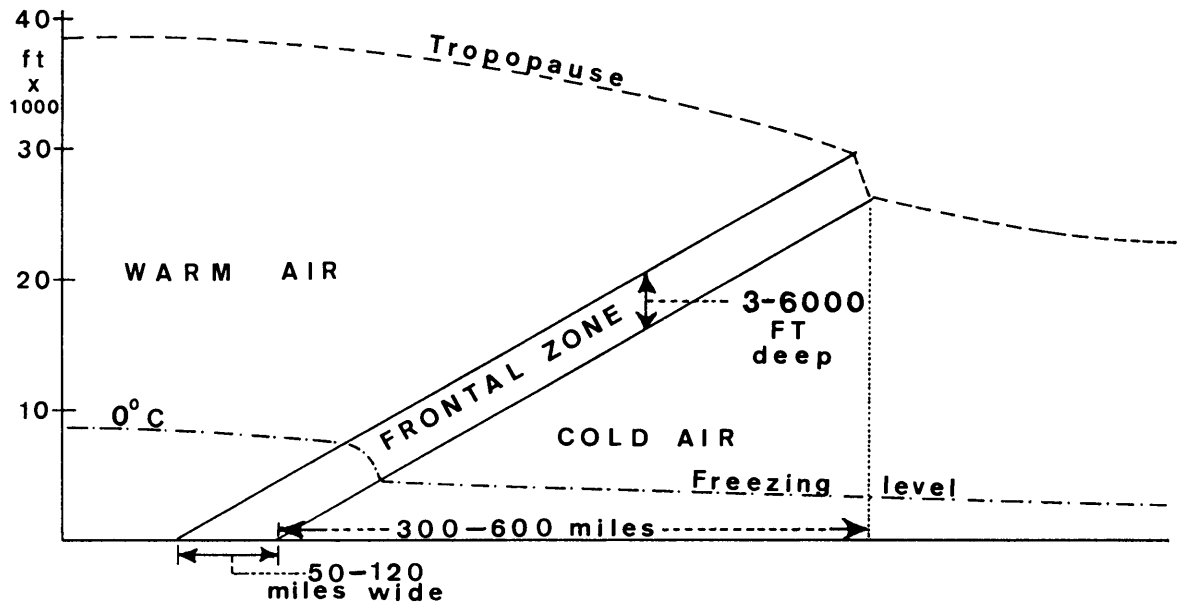
Lows become much deeper when the rising air carries a lot of moisture in it. This is one reason why depressions which draw in warm, moist air from the tropics can become much larger and more intense than those which develop in dry desert regions, or in cold arctic air where there is very little moisture.

Life history of a frontal depression

A large number of depressions develop on an old front where there is already a marked temperature contrast over a relatively short distance. The sequence of events is shown in fig. 23.

On the left-hand side the sequence from (a) to (e) shows how the pattern changes at the surface and up

21. How convergence can produce fronts by concentrating isotherms



22. Cross-section of a front

aloft. The right-hand side shows the same sequence viewed from a satellite.

The sequence is drawn for the northern hemisphere. Developments in the southern hemisphere are a mirror-image of this. Isobars are shown by full lines; pecked lines show the 300 mbar contours which represent the flow at about 30,000 feet. The band of strongest winds is shaded in and marked 'JET'. These two levels have been combined to illustrate the interaction between high- and low-level flow in the development of a depression.

(a) shows the initial stage when the front is quasi-stationary (hardly moving). The surface isobars are very widely spaced indicating light winds. Up aloft there are strong winds blowing almost parallel to the front. Apart from a slight curve at the left the upper contours run almost straight at this stage. The strongest part, the jet, runs to the north of the surface front. A small dotted circle near the right-hand entrance to the jet marks the region where pressure begins to fall.

(b) is the next stage: a depression has formed and deepened enough for the isobars to show a closed circulation of air at low level round the centre. This circulation has begun to twist the line of the front. At this stage the system is called a 'frontal wave' or a 'cold front wave'.

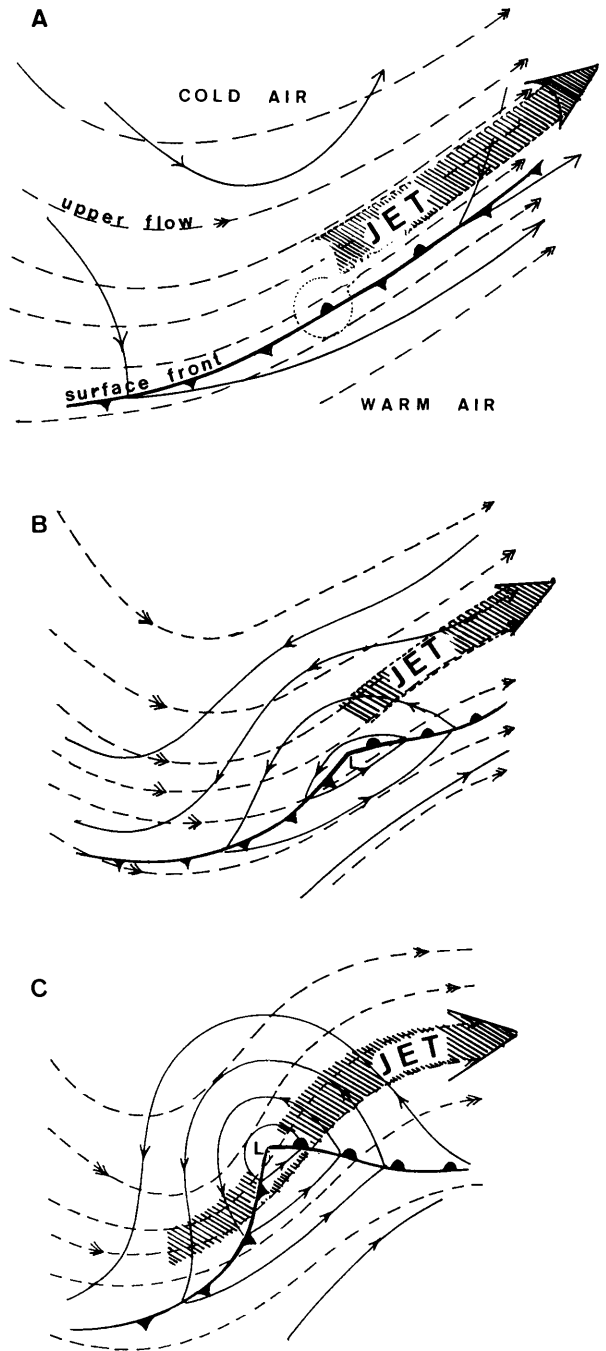
On the western side of the new low the flow has become northerly; this takes colder air southwards. The upper flow responds to this by developing a more marked trough west of the surface centre. The increased curvature round this upper trough increases the upper divergence where the winds accelerate away from it. This in turn increases the pressure fall at the surface (another example of positive feedback).

At (c) the surface low has become fairly deep and the front has been twisted into an inverted 'V' shape. The inside of the V is called the 'warm sector' because it contains the warmest air. On the eastern side, where the cold air is being replaced by warm air, the front is called a 'warm front' and has been identified by rounded blobs. On the western side, where warm air is being replaced by cold, it is called the 'cold front' (identified on charts by triangular spikes).

At high levels the flow has become much more curved, with a well-marked 'upper trough' in the contours and an S-bend in the jet stream. This is usually the stage when the depression deepens most rapidly.

(d) shows that part of the warm front has been

overtaken by the following cold front to form an 'occlusion'. There is still a large warm sector but it is separated from the centre of the low. This is an important step because a significant part of the energy needed to form this low came from the latent heat released when the warm air was lifted to form cloud.

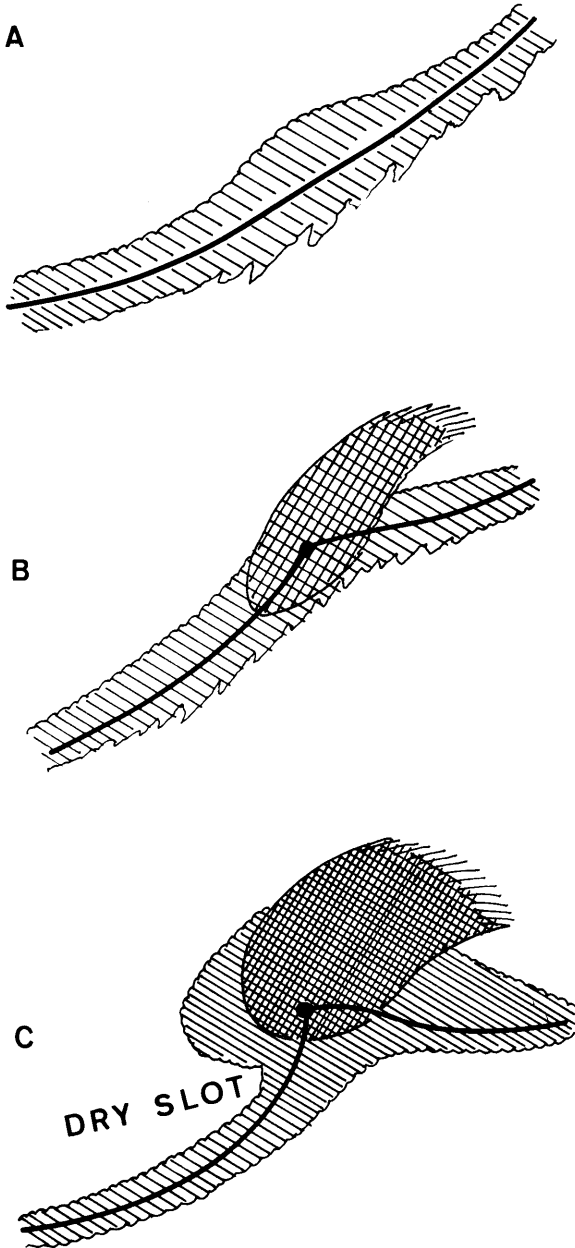


Another important change has taken place aloft. The axis of the jet stream has been transferred and now lies across the tip of the warm sector. The upper trough to the west has become so deep that a small circulation has formed south-west of the surface low. In the early stages the circulation was confined to a shallow layer near the surface. As the depression

deepened the circulation extended higher up. At this stage it has reached 30,000 feet but later it may extend into the lower reaches of the stratosphere.

These two stages, occlusion (leading to separation of warm air from the centre of the low), and the growth of a cyclonic circulation up to high levels, mark the end of the most active period. From now on the main activity is transferred to the outer regions where there is still a supply of warm air and the jet stream is active.

(e) is the final stage: a long occlusion is wrapped round the centre which is filled with cold air. There is no warm air for hundreds of miles. The cyclonic circulation is firmly established up to the base of the stratosphere and there is now very little to displace the centre. The jet stream is far away but still powerful. A little to the south of the jet the cold front has come to a halt and a new wave is beginning to develop. This is the start of another cycle.



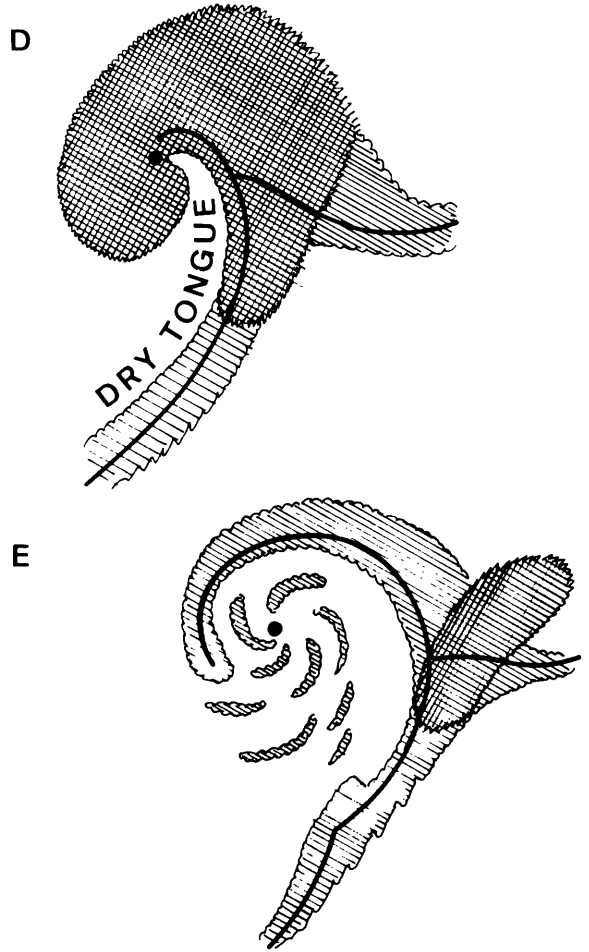
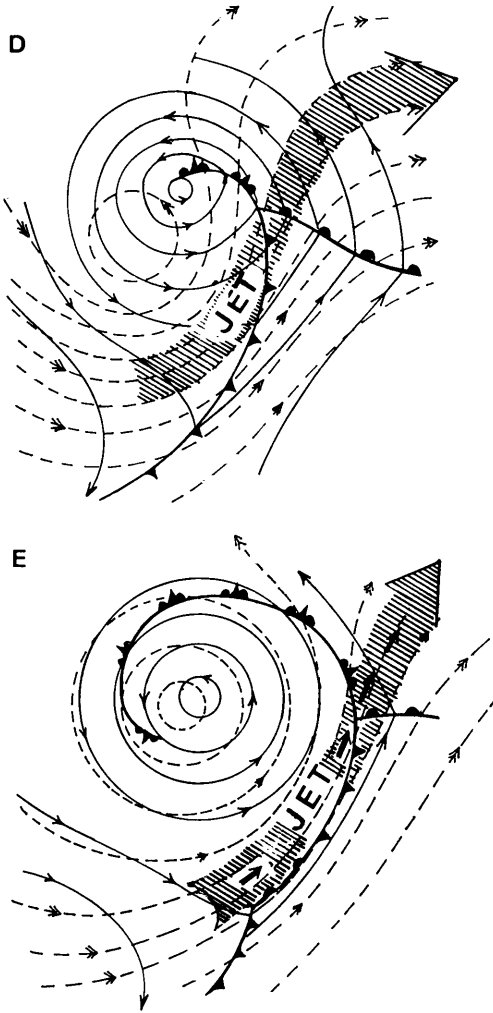
A satellite view of the developing low

The right-hand side of fig. 23 shows how the development appears on satellite pictures. The fronts are marked by thicker black lines. Lower clouds are shown with single hatching, thicker clouds whose tops reach much higher are double hatched.

(a) The quasi-stationary front is marked by a long but fairly narrow band of cloud. In one section the band is thickened by a small bulge on the northern side. This is the first sign of wave development and it often appears on satellite pictures before the surface charts show any sign of a new low.

(b) Now the cloud over the bulge has grown deeper; the cross-hatched region shows where the main ascent of air has taken place. A wedge of high cloud is drawn out into the jet stream to form a finger pointing almost parallel to the front. On infra-red pictures, which emphasise any temperature contrasts, this cloud appears as a bright, white streak showing that it is very cold (and therefore very high).

(c) This is the stage of maximum deepening when the warm air is rising over a wide area to form a shield of cirrostratus ahead of the depression. On high-resolution pictures one may see that the cirrus edge consists of filaments which fan outwards suggesting that the flow aloft is diverging. When such fanning out appears it is a reliable sign of marked deepening at the surface.



23. *Life history of a frontal depression shown by charts and satellite pictures*

The western flank of the cloud mass has a small indentation. This is called the 'dry slot'. It marks the incursion of very dry air which has descended from high levels, sometimes from the stratosphere. Its appearance is the first indication of the separation which will develop between the central part of the low and the warm, moist air to the south.

(d) This is the appearance when the depression is near its deepest stage. Occlusion has taken place and the warm air is being lifted clear of the surface. A long, clear region known as the 'dry tongue' marks the spread of dry air curving round the southern side of the low. There is a large mass of thick cloud but much of it is wrapped round the centre instead of being blown out ahead. This wrap-around indicates that the circulation is no longer confined to the lower levels but now extends up to cirrus levels as well.

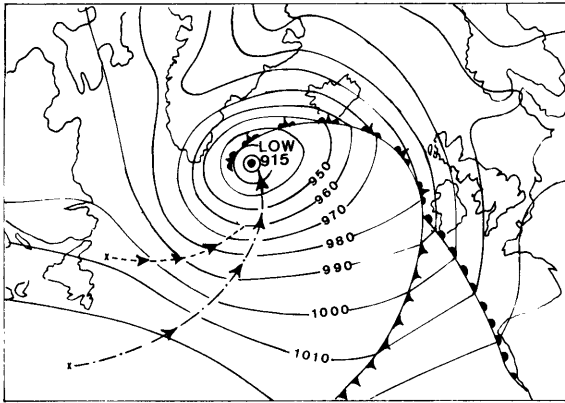
(e) In the final stage there is little left of the warm air near the centre. The occlusion is marked by a thinning band of cloud curving round the centre. Closer in, where the air is both cold and unstable, there are arcs of cloud. These are curving lines of Cu and Cb which rotate round the centre. Some depressions are marked by almost continuous lines curving into the centre like a spiral nebula on an astronomical photograph.

Explosive deepening: 'bombs'

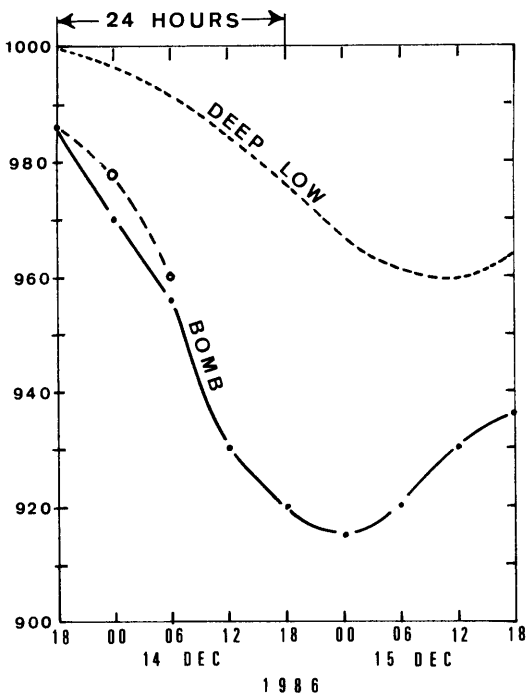
When one large depression becomes slow moving and starts to fill up, secondary lows often form on its perimeter. These secondaries usually start as waves on the cold front. The waves then deepen, grow larger

and become major lows. The sequence may be repeated a number of times and the whole series is called a 'family of depressions'.

Occasionally one member of this family starts to deepen with exceptional rapidity. The Americans have called them 'bombs'. The central pressure of a bomb falls at more than 1 mbar per hour for at least 12 hours. A shallow low starting off at about 1,010 mbar would be down to 960 mbar within 2 days.



24. The exceptionally intense depression of December 1986



25. How central pressure fell in a normal deep low and in a 'bomb'

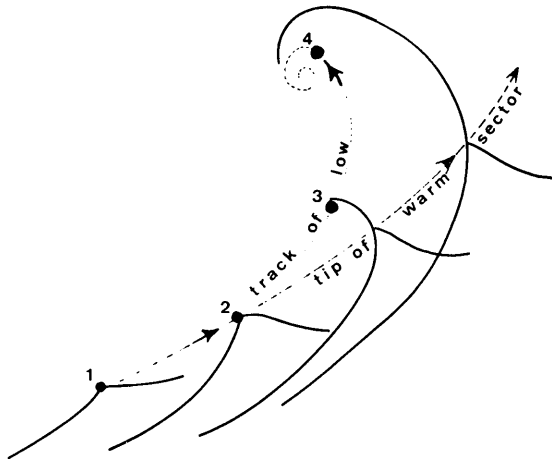
The deepest Atlantic low recorded in recent years started as quite an ordinary wave of 986 mbar south of Newfoundland at 1800 hrs on 13 December 1986. Thirty hours later it had deepened by at least 70 mbar and was centred between Iceland and the southern tip of Greenland. The exact pressure at the centre is not known; the UK analysts drew it as 916 mbar but the West Germans suggested 912 or 913 mbar. Figure 24 shows the storm at its most intense when the circulation covered a large part of the North Atlantic. Figure 25 shows how the central pressure dropped and compares the bomb with a more normal deep depression.

The atmosphere needs to be wound up fairly tightly before a new low turns into a bomb. The first of a family of lows does not deepen so dramatically. It merely sets the stage for its much fiercer offspring. In the December 1986 example the wave developed over the Gulf Stream where it had a supply of warm, moist air below and a jet stream above. There was an extra factor on this occasion—another secondary low following a converging track to the north. The tracks met, the lows merged and then came the period of exceptional deepening. For a time the central pressure fell at more than 4 mbar per hour. (Note: barographs show that in any one place the pressure may change even more rapidly; this is because the instrument responds both to the central deepening and the motion of the low.)

Fortunately most bombs reach their peak intensity out over the oceans, far away from the centres of population. When they do cross land the result can be disastrous. During the afternoon of 15 October 1987, two secondary lows merged over the Bay of Biscay. The combination deepened rapidly to about 958 mbar and wound up into an intense depression which moved across central England during the early hours of 16 October. In a sector to the south and east of the centre, surface winds gusted to 100 knots or more over Brittany and the Cherbourg Peninsula and reached 90 knots near the south coast of England. There was a lot of structural damage and millions of trees were blown down.

The track of depressions

The flow aloft is usually the most important factor controlling the movement of a depression. Figure 26 shows the kind of track followed by the system



26. *Track of a depression and movement of fronts*

illustrated in fig. 23. The way that the track is influenced by the flow aloft is shown in fig. 27. Surface isobars are full lines and the upper flow is represented by pecked lines.

In (a) the surface low is only a shallow feature and the upper flow is a straight west to east current. At this stage the new low travels east.

In (b) the deepening of the surface low has begun to twist the upper flow bringing the high level winds back to a south-westerly direction and steering the low north-east.

By stage (c) the upper trough has deepened so much that a closed centre has formed in the contours aloft. This swings the surface low round on to a northerly track.

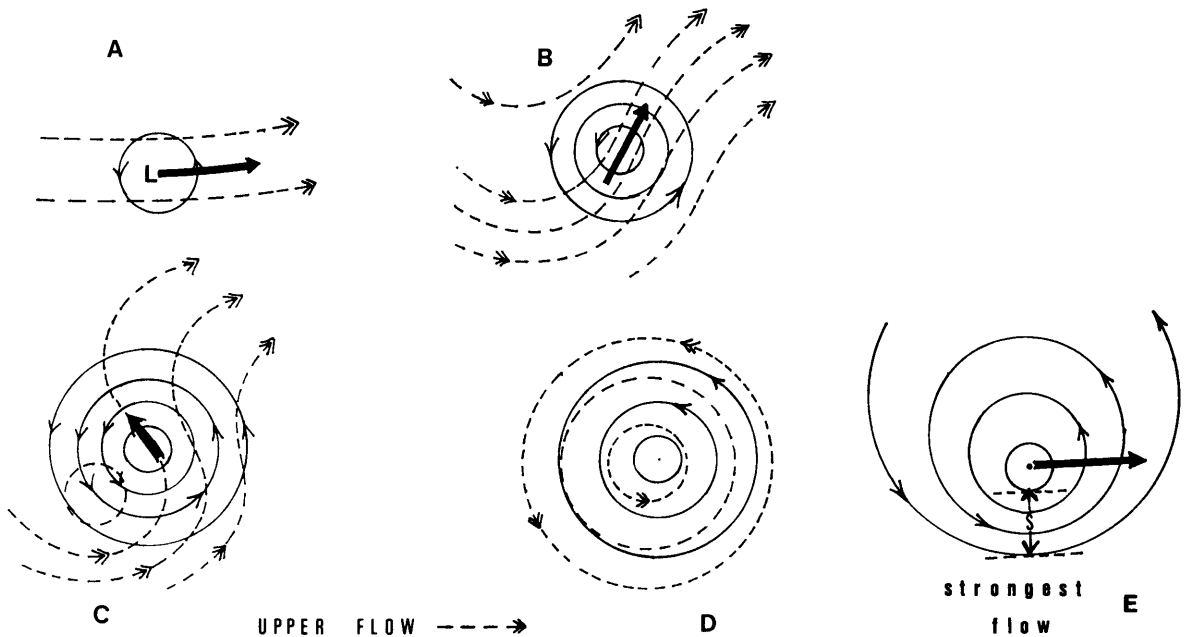
The final stage (d) has an almost solid circulation extending from the surface up to the base of the stratosphere. There is little to steer the old low and the centre becomes very slow moving, often describing tiny circles like the wobbles of a top running down.

Old lows like this do start to move again when the system becomes asymmetrical (with light winds on one side and strong winds on another). The centre then moves in the direction of the strongest winds. This is shown in (e) where the strongest winds are on the south side (blowing from a westerly direction). The centre then moves east.

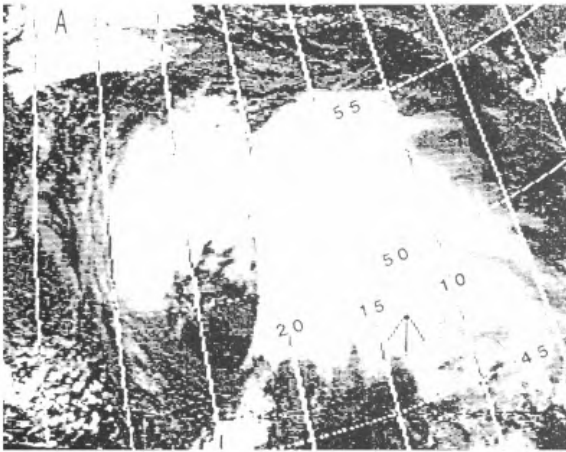
An example of a deep low from peak activity to decline

Figure 28 (a–e) illustrates a series of five infra-red pictures from NOAA-9 and NOAA-10 received between 31 January and 1 February 1988. Cold, high clouds appear very white, lower clouds are grey, the sea (being relatively warm) appears black. Lines of latitude and longitude are marked every 5 degrees.

Picture (a) on the morning of 31 January shows a

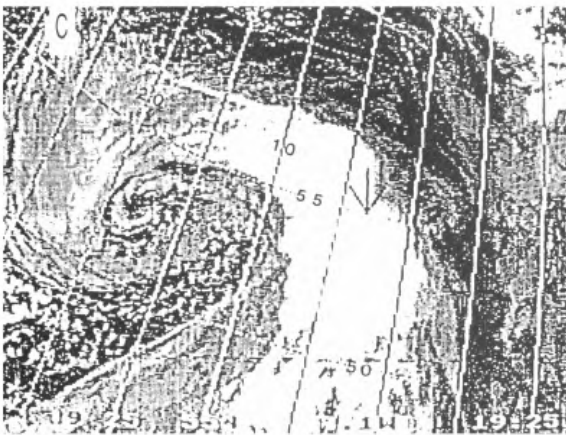


27. *How upper flow controls track of surface depression*

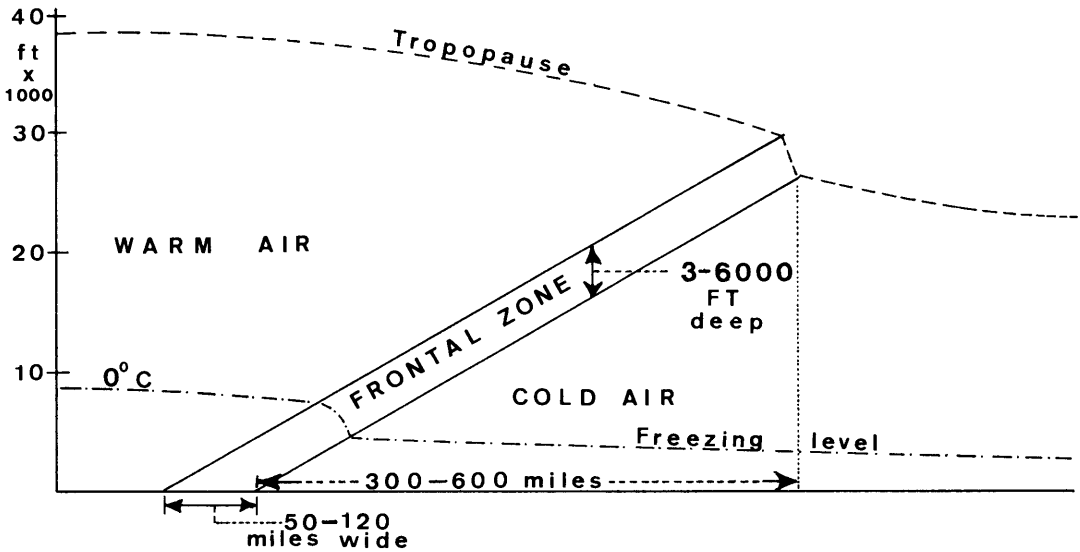


similar stage to fig. 23 (d) when the occlusion was already well developed. Pictures (b) and (c), for the afternoon and evening, show the swirl of cloud winding up round the centre of the depression. Notice that the highest cloud (shown white) begins to disappear where the occlusion is being wrapped around the low. Pictures (d) and (e), taken in the early morning and mid-afternoon of 1 February, illustrate the decaying stages. The curved band of frontal cloud has moved well out from the centre which is marked by curved segments of Cb clouds.

28. (a) to (e) Satellite views of a depression from peak activity to decay



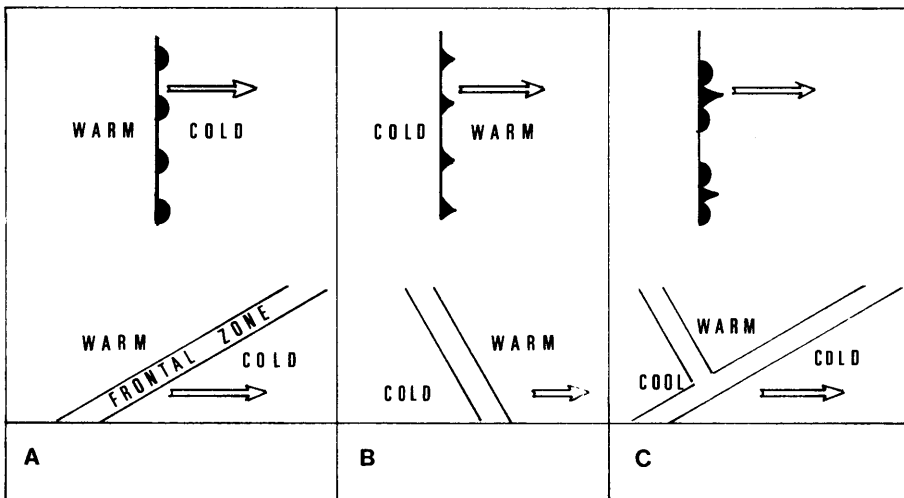
3 Fronts



29. Cross-section of a frontal zone

A front is the boundary between two air masses of different density. The frontal surface usually has a slope of about 1 : 100 with the warmer (less dense) air lying above the colder air.

Figure 29 shows a cross-section of a front. The tropopause (which marks the base of the stratosphere) is shown by a pecked line with a kink where the front extends up to it. On some occasions the



30. (a), (b) and (c) How fronts are marked on charts and what their cross-sections look like

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