Winds in the atmosphere

Measuring the wind; the Beaufort scale

The wind is one of the more obvious features of the weather. While we have a rough and ready appreciation of what counts as a ‘strong’ or ‘gentle’ wind, this does not give any quantitative estimate. By about 1800, it was clear that it would be useful to categorise winds in terms of their effects on sailing ships. This was done by Francis Beaufort, whose scale remained standard until 1946 and is still that used for shipping forecasts. The table below gives a simplified modern description for use on land.

Beaufort Number	Description	Wind speed (m.p.h.)	Wind speed (m/sec)
0	*Calm* ; smoke rises vertically	0	0-0.5
1	*Light air* ; wind direction shown by smoke drift, not by vanes)	1-3	0.5-1.5
2	*Light breeze*; wind felt on face, leaves rustle, vanes move	4-7	1.5-3.5
3	*Gentle breeze:* leaves and small twigs move, light flags lift, large wavelets at sea	8-12	3.5-5.5
4	*Moderate breeze;* dust and loose paper lift, small branches move	13-18	5.5-8.0
5	*Fresh breeze:* small leafy trees sway, moderate waves	19-24	8.0-10.5
6	*Strong breeze:* large branches sway, telegraph wires whistle, umbrellas difficult to use	25-31	10.5-13.5
7	*Near gale;* whole trees move, inconvenient to walk against	32-38	13.5-17
8	*Gale;* small twigs break off, walking impeded, high waves and foam	39-46	17-20.5
9	*Strong gale;* slight structural damage	47-54	20.5-24.5
10	*Storm;* considerable structural damage, trees uprooted	55-63	24.5-28.5
11	*Violent storm;* widespread damage	64-72	28.5-32.5
12	*Hurricane;* at sea, visibility badly affected by driving foam and spray; sea surface completely white	>73	>32.5

***Where do the winds come from* ?**

The atmosphere is (in one sense at least) a gigantic heat engine. The radiation from the sun causes convection both on a local scale (as we saw when considering the energy balance in the section on solar radiation) and on a global scale. The difference in energy delivered at the equator and the poles creates the pressure differences that drive the major wind systems in the earth’s atmosphere. Winds are masses of air in motion. An *air mass* has a reasonably precise meaning in meteorology. This is a large volume of air (covering millions of square kilometres) that has reasonably constant pressure and humidity. Thus the air mass will determine the overall weather of a region (although not the local microclimate). They come from the extensive regions of high pressure that characterise some of the areas of the earth; the subtropical oceans (throughout the year) and the mid and high latitude continents (mainly in winter). Air spirals out from these high-pressure areas (as we shall see they are *anticyclones*) to create the wind systems of the planet. Two examples of relevance to Europe are

- *the Azores anticyclone*. The south westerlies blow towards the North Pole and the North-East Trades towards the equator. This warm and humid air is classified in mid-latitudes as 'tropical maritime'
- *the polar continental*; the cold dry air that comes from the Eurasian continent in winter.

The weather in the British Isles changes frequently since air masses arrive from different regions. The boundaries that separate air masses are called *fronts* (the name comes from the trenches of the 1st World War). Hereabouts a *cold front* is the leading edge of a cold air mass and brings rain. The warm moist air of the tropical maritime mass it is replacing is forced upwards, cools and water precipitates out.

The principal forces acting on a parcel of air

If we are to understand why the winds occur, it is necessary to consider the forces that act on the air masses in the atmosphere. To any observer stationary with respect to the surface of the earth, there are four forces acting on a parcel of air in the atmosphere

- gravitational
- pressure gradient
- Coriolis force
- frictional force

We will consider each of these in turn

The gravitational force

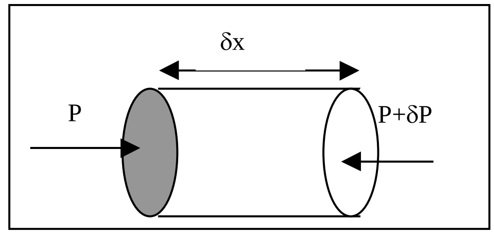
Due to the large mass of the earth, the gravitational force is one of the strongest forces acting on the air parcel and is directed towards the centre of the earth

$$F_g = g\rho\Delta V \qquad (1)$$

where g is the gravitational constant (more or less constant through the troposphere), ρ is the density of air and ΔV is the volume of the parcel. The parcel also experiences a centrifugal force due to the rotation of the earth. This is much smaller than the gravitational force, but should still be allowed for in accurate calculations. The usual way is to replace g by an effective gravitational constant g'. This is smaller than g since the centrifugal force acts in the opposite direction to the gravitational force.

The pressure gradient

The pressure at the surface of an air parcel is the normal component of the force exerted by its surroundings on a unit area of surface. (Remember that the definition of pressure is force per unit area). This force is always directed towards the parcel. The parcel will experience a net force if there is a difference between the pressures on the surfaces at different sides.



Consider the diagram on right where the shaded cross-section has area δA. The net force on the parcel due to the pressure difference is

$$F_P = P\delta A. - (P + \delta P)\,\delta A = -\delta P\delta A \qquad (2)$$

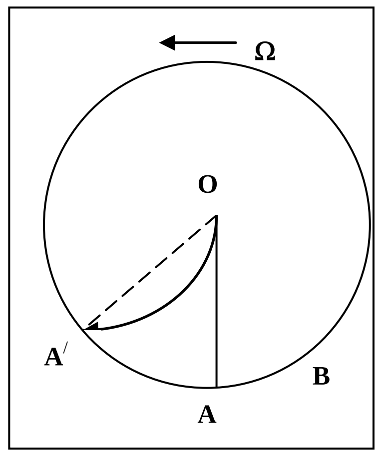
If ρ is the density of air, then the force per unit mass (i.e. the acceleration) is given by

$$\frac{F_P}{\rho\delta A\delta x} = -\frac{1}{\rho}\frac{\delta P}{\delta A}\frac{\delta A}{\delta x} = -\frac{1}{\rho}\frac{\delta P}{\delta x} \qquad (2)$$

taking the limit in the usual way and generalising to three dimensions, we get

$$\mathbf{F}_P = -\frac{1}{\rho}\nabla P \qquad (3)$$

The Coriolis force



Only the gravitational and pressure gradient forces can initiate motion of the air but there is a third, 'fictitious force' due to the rotation of the earth – this is the Coriolis force. Consider the diagram on the right which represents a region around the North Pole (O) such that we can consider it to be a rotating disc. An air parcel starts to move horizontally away from the pole towards a point A. If no forces act on this parcel, by Newton's laws it will follow a direct path OA. However, the disc is also rotating with an angular velocity Ω, and so it will follow the curved line OA/ with respect to the disc. The disc has moved on, and so the point A is at B by the time the parcel of air reached the edge. To an observer rotating with the earth, it looks as though the parcel is deflected by a force away from A towards A/ The fictitious force is called the Coriolis force. (A more systematic way of looking at this problem is that the simple application of Newton's Laws of the form $F = ma$ is not applicable in rotating frames of reference, however, in a case like the earth it is much more convenient to take the earth as stationary and consider the matter in terms of fictitious forces.) The earth is a sphere and not a disc, which means that we should use the full vector notation, but the principles are the same. We must calculate the vector product between the rotation vector $\mathbf{\Omega}$ (directly out of the plane for the disc) and the velocity vector, $\boldsymbol{V_g}$

Applying this to the earth, if we consider an air parcel with velocity $\boldsymbol{V_g}$, and the angular velocity vector of the earth is $\mathbf{\Omega}$, then the Coriolis force per unit mass is given by $\boldsymbol{f} = -2\mathbf{\Omega}\times\boldsymbol{V_g}$. However, since the atmosphere is thin compared with the radius of the earth, we know that the wind is blowing almost along the *local* horizontal. Thus it is useful to split the Coriolis force into two contributions with respect to the *local* vertical. We designate these by the unit vectors **z** and **y.** If the latitude is φ, then the two components are given by $(\Omega\sin\varphi)\mathbf{z}$ and $(\Omega\cos\varphi)\mathbf{y}$ where **z** is the unit vector along the local vertical, **y** is in the local horizontal plane. If we assume that the wind is moving in the **y** direction, then the Coriolis force is given by

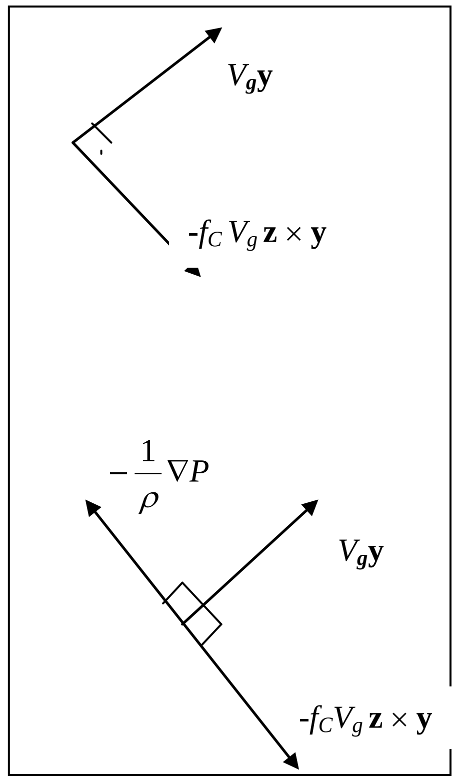
$$\mathbf{f} = -(\Omega \sin \varphi)\mathbf{z} \times V_g \mathbf{y} = (V_g \Omega \sin \varphi)\mathbf{x} \qquad (4)$$

i.e. there is a force in the **x** direction. To an earth- bound observer, the wind is steered increasingly to the right of the initial direction of movement when viewed downward in the northern hemisphere and increasingly to the left in the southern hemisphere. The initial movement will be down the pressure gradient (which is providing the force that creates the wind), but deflections continue until the wind is blowing perpendicularly to its original motion (and so at right-angles to the pressure gradient producing it). It cannot then be deflected any further; the Coriolis force *balances* the pressure gradient. Put formally

$$\left(-\frac{1}{\rho}\nabla P + f_C V_g\right)\mathbf{x} = 0\,; \quad f_C = 2\Omega \sin \varphi \qquad (5)$$

where f_C is the *Coriolis constant* (about 10^{-4} sec^{-1}), V_g is the air velocity at balance. This gives

$$V_g = \frac{1}{f_C \rho}\frac{dP}{dx} \qquad (6)$$



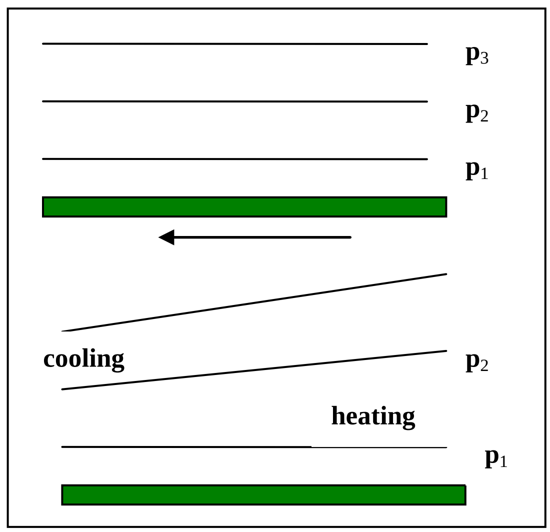
assuming that the pressure gradient increases along the **x** axis. The diagrams on the right show the effect. The point where the Coriolis effect balances the pressure gradient is called the *geostrophic balance* and the resulting wind called the *geostrophic wind*. The important point to note is that the wind direction is parallel to the isobars and (in the northern hemisphere) the wind direction is such that the lower pressure is on the left-hand side as you face downwind. Thus a *low-pressure* area in the northern hemisphere has the winds rotating around it in a *counter-clockwise* direction (but clockwise in the southern hemisphere). This type of motion is called *cyclonic* . Thus low pressure weather system is called a cyclone. Over a high-pressure area in the northern hemisphere, the geostrophic wind circulates in a clock-wise direction. This is called *anticyclonic* motion and therefore a high-pressure weather system is known as an anticyclone.

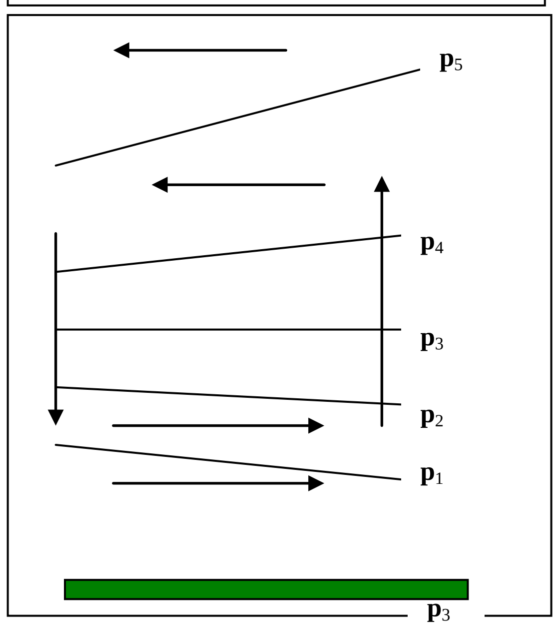
From the value of f_C derived above, it is clear that the Coriolis force is greatest at the poles, and decreases as one approaches the equator, where it is zero. (Note that there will also be a vertical component proportional to cos φ arising from the **y** vector; this will make a small (negligible in fact) contribution to the effective gravitational force).

The frictional force

There exists considerable frictional force between the atmosphere and the earth's surface (for example due to mountains and hills. Lines of tall trees are planted to act as wind breaks and protect crops). Frictional forces are difficult to treat properly. The mechanism is essentially a form of viscosity (at low altitudes) and small-scale eddy mixing processes at higher altitudes. The layer where the frictional force is important is known as the *planetary boundary layer*. The thickness of the layer is very variable; from a few hundred metres in still air at night to 4-5km over a hot surface with strong convection.

This introduces the question of how the wind speed varies with height which is obviously closely tied to the question of how atmospheric pressure varies with height. We considered this question when we looked at the structure of the atmosphere. Its connection to the wind speed can be seen through the following example.



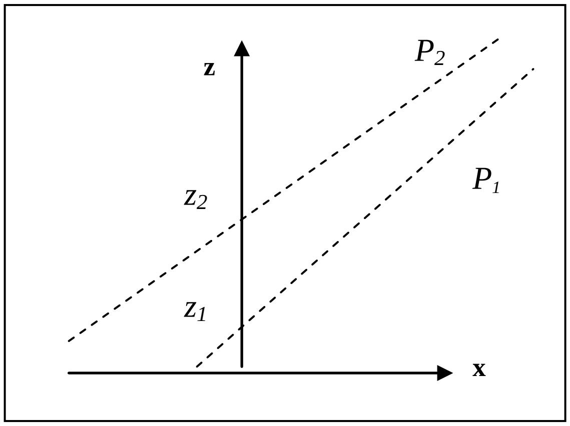


Thermal gradients and winds

Pressure differences can be induced by uneven heating or cooling. Consider a region of the atmosphere with an even temperature distribution (top right with $p_1 > p_2 > p_3$). The air is now heated at one end (adding energy) and cooled at the other (extracting energy). Hot air expands and cold air contracts. Thus the air columns on the left and right hand sides will try to expand (on the right) and contract (on the left) This will produce a horizontal pressure gradient; the lines joining points of equal pressure (*isobars*) will no longer be at the same height. There is therefore a force moving air down the pressure gradient (bottom right). This is a *thermal wind.* The rate of decrease of pressure at a fixed height in the warm area is equal to the mass of air flowing out above the fixed height. The rate of pressure decrease is therefore larger in the lower levels than the higher. Eventually a steady state is reached. The pressure in the *upper part* of the air column will be *lower* in the *warm* area and *higher* in the *cold* areas; whereas in the *lower part* of the air column the pressure is *higher* in the *cold* areas and *lower* in the *warm*. This is the basis of *convection* in the atmosphere.

Example: a day at the seaside: During the day, the land is heated by the sun and the temperature rises above that of the sea (the specific heat of the land is less than the specific heat of the sea). The air over the landis warmer than that over the sea resulting in air blowing from the sea towards the land (at low levels) – a sea breeze. At night the land cools below the temperature of the sea and so the situation reverses. The (low-level) wind now blows from land to sea – a land breeze. This simple example shows one of the basic mechanisms underlying the global circulation of the atmosphere.

We can put this argument on a more formal basis and consider the variation of the geostrophic wind with height. Let us consider this in more detail. For simplicity, let us assume as before that the pressure variation is along the (+**x**) axis. Then the geostrophic wind is along the (+**y**) axis and is given by



$$V_g = \frac{1}{f_C \rho}\frac{dP}{dx}$$

We now consider the situation in the diagram on the right. We take two isobars, pressure P_1 and P_2 at heights z_1 and z_2 and calculate the difference in the geostrophic wind velocity between them in the upwards direction. This is given by

$$\Delta V_g = \frac{1}{f_C}\left[\left(\frac{1}{\rho}\frac{dP}{dx}\right)_{z_2} - \left(\frac{1}{\rho}\frac{dP}{dx}\right)_{z_1}\right] \qquad (6)$$

But we know that the hydrostatic equation states that

$$dP = -\rho g dz \qquad (7)$$

In the diagram above, the isobars are sloped, so we can measure dP either along the **x** axis or along the **z** axis. In the diagram above, the values of z are linked to the lines of constant pressure that intersect the **x** axis, so we can write

$$\frac{dP}{dx} = -\rho g \frac{dz(P)}{dx} \qquad (8)$$

where we have written $z = z(P)$ to remind us of the connection with the heights and the isobars. So we can therefore write (6) as

$$\Delta V_g = \frac{g}{f_C}\left[\left(\frac{dz(P)}{dx}\right)_{z_2} - \left(\frac{dz(P)}{dx}\right)_{z_1}\right] = \frac{g}{f_C}\frac{d(\Delta z)}{dx} \qquad (9)$$

We can now use the buoyancy equation, $dP = -\rho g dz$, again to determine Δz. Using the ideal gas law, $P = R_S \rho T$ and rearranging, we have $dP = \frac{Pg}{R_S T} dz$. Integrating between the isobars P_1 and P_2 (and therefore between the heights z_1 and z_2) we have

$$-\int_{P_1}^{P_2} \frac{R_S T}{g} \frac{dP}{P} = \int_{z_1}^{z_2} dz \tag{10}$$

where R_S is the specific gas constant. Integrating this and assuming that g is constant, we obtain

$$-\frac{R_S}{g} \int_{P_1}^{P_2} T(z) d \ln P = z_2 - z_1 \tag{11}$$

If we replace the height-dependent temperature $T(z)$ by an average value, $\overline{T}$, then we obtain

$$z_2 - z_1 = \Delta z = \frac{R_S \overline{T}}{g} \ln\left(\frac{P_1}{P_2}\right) \tag{12}$$

We can now differentiate this, remembering that we chose P_1 and P_2 to be constant.

$$\frac{d(\Delta z)}{dx} = \frac{R_S}{g} \ln\left(\frac{P_1}{P_2}\right) \frac{d\overline{T}}{dx} \tag{13}$$

Substituting this into (9) we get

$$\Delta V_g = \frac{R_S}{f_C} \ln\left(\frac{P_1}{P_2}\right) \frac{d\overline{T}}{dx} \tag{14}$$

Finally, we can eliminate the pressure term using (12) to give the difference form of the thermal wind equation

$$\frac{\Delta V_g}{\Delta z} = \frac{g}{f_c \overline{T}} \left(\frac{d\overline{T}}{dx}\right) \tag{15}$$

As ΔV_g and Δz tend to zero, the left-hand side tends to the derivative and the average temperature on the right-hand side tends to that of the isobaric surface in the middle of the layer. This gives an important relationship between the *vertical* variation of the geostrophic wind and the *horizontal* temperature gradient. The isobaric slope increases with height, so the horizontal pressure gradient also increases with height. Hence the geostrophic wind should also increase with height. The increase in velocity is in a direction which is perpendicular to the temperature gradient with the *cold region* to the left and the *warm region* to the right of the wind vector increment (in the northern hemisphere). (In our case the temperature gradient is in the +**x** direction and the increase in wind velocity is in the +**z** direction). Thus the geostrophic wind rotates counter-clockwise (with altitude) when the wind blows from a cold region to a warm one, and clockwise (with altitude) when the wind blows from warm to cold.

Cyclones and anticyclones

Storms have been recorded throughout history, but the first comprehensive account of a storm is given by Daniel Defoe (the 'great storm' of 26th November 1703). The first observation on cyclonic systems is that of Benjamin Franklin (21st October1743). He noted that he was prevented from seeing a lunar eclipse in Philadelphia by a N.E. gale *but* the eclipse was seen in Boston (300km to the N.E.) Thus the storm itself must be moving *against* the direction of its constituent winds. The development of the electric telegraph allowed storms to be tracked. Admiral Fitzroy founded the U.K. Meteorological Office. He noted that cold and hot air masses were both involved in cyclonic behaviour (1840s). From 1910 to 1930, the Norwegian Meteorological Office plotted warm and cold fronts – the first use of scientific aircraft. After 1960, the use of meteorological satellites grew rapidly, enabling the plotting of cyclonic motion and 'fronts'.

Cyclonic systems can be found in low and middle latitudes. Middle latitude cyclones are the systems responsible for the 'bad weather' in middle latitudes. They are usually called *depressions* (not because of this but because they are *low pressure* systems). Cyclones that arrive here form on the other side of the Atlantic (where cold polar air over the North American continent meets warm tropical air from the Western Atlantic – heated by the Gulf Stream). Similar effects occur in the North Pacific and Northern Mediterranean (alpine air meeting air from North Africa). A depression forms when a wave develops at the boundary between the two air masses. Air then starts to flow across the isobars and a low pressure area develops with cyclonic motion. This then moves in accordance with the winds in the warm sector. Since cold air tends to move faster than warm air, the cold front tends to catch up with the warm front and the warm air rises above the cold air.

Anticyclones form over Siberia, Canada and North Russia. These dominate the Asian and North American climate in winter. They give rise to (mainly) dry weather but can trap extensive low-lying clouds. Since winds are usually light, pollution levels often rise. Most anti-cyclones last only four or five days, but occasionally they last much longer. Such systems are called *blocking highs.* These block the movement of cyclones in mid-latitudes, giving stationary fronts for days and even weeks. (building up smoke, pollution etc. two notable examples in the U.K. were Jan-Mar 1963 and July-Sept 1976).

Tropical cyclones (also known as *hurricanes* [probably from a Taino (Caribbean) word *hurakan* – a storm god] or *typhoons* [either Urdu *tu fan* or perhaps dialect Chinese *dai fung*]). These are low pressure systems of great intensity. Wind speeds of 100m/sec (225 miles/hour) can be produced and heavy rainfall always occurs. The cyclonic nature of air circulation in the lower atmosphere in a hurricane or typhoon is easily seen by the nature and movement of clouds as seen from a satellite. Rising air from the equator is humid and cools on rising giving a girdle of clouds around the earth. These give heavy rain in the equatorial regions. The cold and warm air meet at the *intertropical convergence zone* (ITCZ) and minor depressions along the edge. Some (and only some of these) develop into major storms. The reason for the early stages are not clear, but once begun, the mechanisms that give major hurricanes (or typhoons) is well understood

1. Pressure starts to fall rapidly at the centre of the disturbance
2. Winds *rise* in a tight band, 30-60km in radius (the central eye)

3. As the storm grows it moves to the westerly in the trade winds (8-15^0 latitude) and migrates to higher latitudes
4. The maturing storm expands while the central pressure stops falling. The route depends on local surface conditions such as the surface temperature (and perhaps the salinity) of the sea (and is unpredictable). Hurricanes only form over warm water (temperature greater than 27^0)
5. The storm grows to 300km radius or greater and then begins to decay. Decay is hastened by passing over cold water or land
6. The storm is pushed by mid-latitude westerlies in higher latitudes as it decays

These storms are found in both hemispheres, most commonly in the North Pacific (where there are few people) . Most frequent in late summer (when the sea surface is warmest) but they can occur all summer and autumn in the tropics. They are part of the process by which energy is transported from the equator to the poles.

Global convection

The first model to describe large-scale global convection was proposed by George Hadley in 1735. He noted that air in the lower latitudes is warmer than that in the higher (polar) latitudes due to the greater solar flux reaching the equator. Tropical air should move (rise) vertically and move northwards while the cool polar air should move southwards. As the tropical air moves north, it loses energy by radiation before descending to the ground, so replacing the southward-moving colder air. Similarly, the cold air will gain heat from the ground (which is itself heated by radiation as discussed earlier) and will therefore rise in the equatorial regions. Thus a circulation system is formed transporting thermal energy from the equator to the poles. This is the *Hadley cell*.

There are, however, significant differences between Hadley's model and the real air circulation patterns. It is true that there is a low pressure belt over the equator and a high pressure region over the pole as the Hadley model would predict. However, there is another circulating cell of air between 30^{0}N and 60^{0}N (and similarly in the southern hemisphere) in which air rises in the colder regions (i.e. at 60^{0}N) and descends in the warmer regions (i.e. in the *opposite* direction to the Hadley mechanism). This is the *Ferrel cell*. Finally, there is a third cell between 60^{0}N (and S) and the pole. This circulates in the same direction as the Hadley cell but is much weaker. It is known as the *Polar cell.*

***Global wind patterns*.**

Before we can describe these, we must discuss how the atmospheric pressure varies over the planet. This depends on the seasons.

- *Northern winter/southern summer*. There is a weak high pressure across the Arctic Ocean and low pressure centres SE of Greenland and in the North Pacific. Between 45^{0}N and 15^{0}N there is an extensive region of high pressure, stretching from the subtropical Eastern Pacific eastwards round the planet as far as India (an exception is a low-pressure region in the Mediterranean). A broad belt of weak pressure gradient encircles the planet at low latitudes. There are weak low pressure centres over the southern continents and subtropical anticyclones over the main southern oceans. At 40^0S there is a belt of rapidly changing pressure gradient (and therefore strong winds) and finally a high pressure region over Antarctica

- *Northern summer/southern winter*. The mid-latitude oceanic low centres in the North Atlantic and North Pacific weaken and the subtropical anticyclones shift towards the Pole. Extensive low pressure (rather than high pressure) is observed over central and southern Asia (associated with the monsoon). There are also low pressure regions over the subtropical areas of Africa and North America. The reverse (from summer low to winter high) occurs over the equivalent areas in the southern hemisphere. There is a deep low pressure belt around Antarctica.

The strongest wind patterns produced are, as one would expect from the seasonal pressure variation discussed above, themselves seasonal

Northern winter/southern summer

- *the mid-latitude westerlies*. (i.e. blowing from the west). In the Northern hemisphere these are confined by the ocean basins.
- *the north-easterlies* moving out from the Asian anticyclone over the Arabian and South China seas
- *the trade winds* – so called because of their importance to sailing ships in the 16^{th} – 19^{th} century. The trade wind in the northern hemisphere is north-easterly (i.e. blows *from* the north-east); in the southern hemisphere it is south-easterly. The *Intertropical Convergence Zone* (*ITCZ*) is the west-east region into which the heat and moisture carried by these winds are blown

Northern summer/southern winter

- *the mid-latitude westerlies* are much weaker in the northern latitudes (about 65% of their winter velocity
- the *ICTZ* over the oceans moves north (but not as far as the continental eqivalent which penetrates into sub-Saharan Africa and southern Asia). This causes a reversal of wind direction over the Arabian and South China seas which is part of the monsoon

Such winds are not explained by Hadley's model but are due to the balance of forces set up by the pressure gradient induced by (i) thermal convection (ii) frictional forces of air moving across the earth's surface and (iii) Coriolis force.

Wind patterns higher up are much simpler. The frictional force of the continental masses is reduced.. The most important seasonal winds occur between 30^0N and 50^0N (and similarly S). These are the *jetstreams.* An important effect in the Southern hemisphere is the *polar vortex*. The westerly winds high over the coasts of Antarctica flow in a more or less circular pattern and prevent heat transport (by incoming depressions) into the air above Antarctica. This increased cold is important in depleting the ozone layer and explains why the 'hole' was first observed over Antarctica.