CHAPTER SIX

Atmospheric motion: principles

6

LEARNING OBJECTIVES

When you have read this chapter you will:

- know the basic laws of horizontal motion in the atmosphere
- know how the Coriolis force arises and its effects
- be able to define the geostrophic wind
- know how friction modifies wind velocity in the boundary layer
- understand the principles of divergence/convergence and vorticity and their roles in atmospheric processes
- understand the thermodynamic, dynamic and topographic factors that lead to distinctive local wind regimes.

The atmosphere is in constant motion on scales ranging from short-lived, local wind gusts to storm systems spanning several thousand kilometers and lasting about a week, and to the more or less constant global-scale wind belts circling the earth. Before considering the global aspects, however, it is important to look at the immediate controls on air motion. The downward-acting gravitational field of the earth sets up the observed decrease of pressure away from the earth's surface that is represented in the vertical distribution of atmospheric mass (see Figure 2.13). This mutual balance between the force of gravity and the vertical pressure gradient is referred to as *hydrostatic equilibrium* (p. 31). This state of balance, together with the general stability of the atmosphere and its shallow depth, greatly limits vertical air motion. Average horizontal wind speeds are of the order of 100 times greater than average vertical movements, although individual exceptions occur – particularly in convective storms.

A LAWS OF HORIZONTAL MOTION

There are four controls on the horizontal movement of air near the earth's surface: the pressure-gradient force, the Coriolis force, centripetal acceleration, and frictional forces. The primary cause of air movement is the development of a horizontal pressure gradient through spatial differences in surface heating and consequent changes in air density and pressure. The fact that such a gradient can persist (rather than being destroyed by air motion towards the low pressure) results from the effect of the earth's rotation in giving rise to the Coriolis force.

1 The pressure-gradient force

The pressure-gradient force has vertical and horizontal components but, as already noted, the vertical component is more or less in balance with the force of gravity. Horizontal differences in pressure arise from thermal heating contrasts or mechanical causes such as mountain barriers and these differences control the horizontal movement of an air mass. The horizontal pressure gradient serves as the motivating force that causes air to move from areas of high pressure towards areas where it is lower, although other forces prevent air from moving directly across the isobars (lines of equal pressure). The pressure-gradient force per unit mass is expressed mathematically as

$$\frac{1}{\rho}\frac{\mathrm{d}p}{\mathrm{d}n}$$

where ρ = air density and $\mathrm{d}p/\mathrm{d}n$ = the horizontal gradient of pressure. Hence the closer the isobar spacing the more intense is the pressure gradient and the greater the wind speed. The pressure-gradient force is also inversely proportional to air density, and this relationship is of particular importance in understanding the behavior of upper winds.

2 The earth's rotational deflective (Coriolis) force

The Coriolis force arises from the fact that the movement of masses over the earth's surface is referenced to a moving coordinate system (i.e., the latitude and longitude grid, which 'rotates' with the earth). The simplest way to visualize how this deflecting force operates is to picture a rotating disc on which moving objects are deflected. Figure 6.1 shows the effect of such a deflective force operating on a mass moving outward from the center of a spinning disc. The body follows a straight path in relation to a fixed frame of reference (for instance, a box that contains the spinning disc), but viewed relative to coordinates rotating with the disc the body swings to the right of its initial line of motion. This effect is readily demonstrated if a pencil line is drawn across a white disc on a rotating turntable. Figure 6.2 illustrates a case where the movement is not from the center of the turntable and the object possesses an initial momentum in relation to its distance from the axis of rotation. Note that the turntable model is not strictly analogous since the outwardly directed centrifugal force is involved. In the case of the rotating earth (with rotating reference coordinates of latitude and longitude), there is apparent deflection of moving objects to the right of their line of motion in the Northern Hemisphere and to the left in the Southern Hemisphere, as viewed by observers on the earth. The idea of a deflective force is credited to the work of French mathematician G.G. Coriolis in the 1830s. The 'force' (per unit mass) is expressed by:

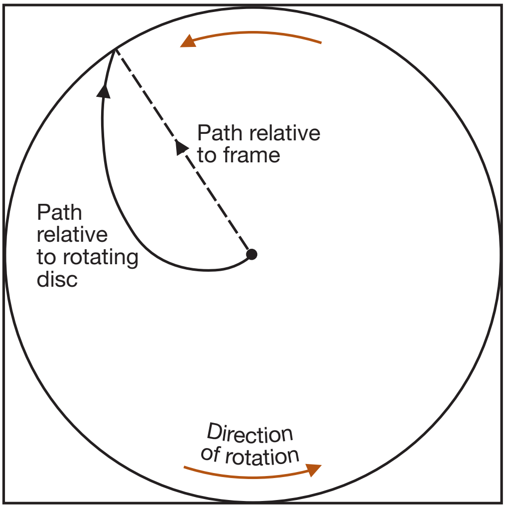

Figure 6.1 The Coriolis deflecting force operating on an object moving outward from the center of a rotating turntable.

$$-2\,\Omega\,V\sin\phi$$

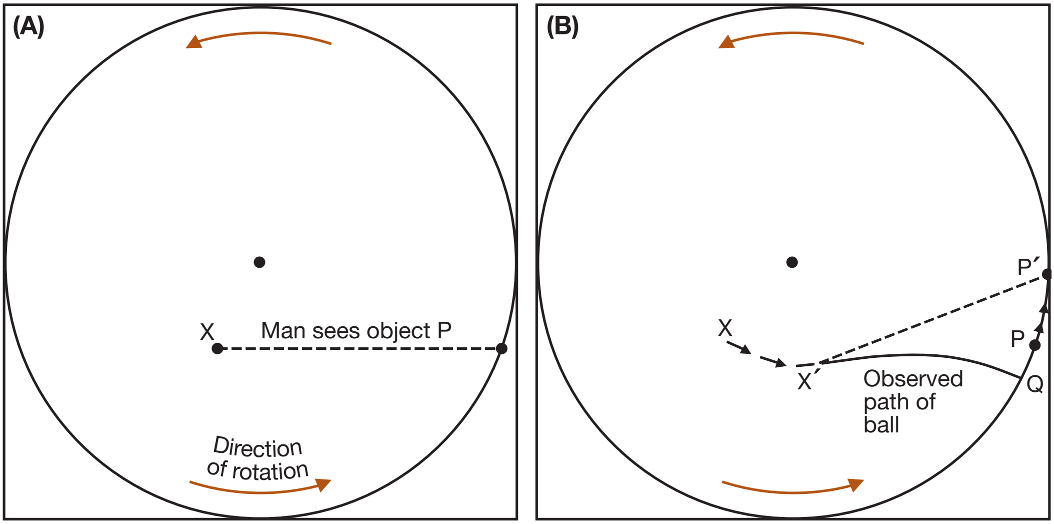

Figure 6.2 The Coriolis deflecting force on a rotating turntable. A. An observer at X sees the object P and attempts to throw a ball towards it. Both locations are rotating anticlockwise. B. the observer's position is now X′ and the object is at P′. To the observer, the ball appears to follow a curved path and lands at Q. The observer overlooked the fact that position P was moving counterclockwise and that the path of the ball would be affected by the initial impulse due to the rotation of point X.

where Ω = the angular velocity (15° hr^{-1} or $2\pi/24$ radians hr^{-1} for the earth = 7.29×10^{-5} radians s^{-1}); ϕ = the latitude and V = the velocity of the mass. $2\Omega \sin \phi$ is referred to as the Coriolis parameter (f). Angular velocity is a vector representing the rate of rotation of an object about the axis of rotation; its magnitude is the time rate of displacement of any point of the body.

The magnitude of the deflection is directly proportional to: (1) the horizontal velocity of the air (i.e., air moving at 10m s^{-1} has half the deflective force operating on it as on that moving at 20m s^{-1}); and (2) the sine of the latitude (sin 0° = 0; sine 90° = 1). The effect is thus a maximum at the poles (i.e., where the plane of the deflecting force is parallel to the earth's surface). It decreases with the sine of the latitude, becoming zero at the equator (i.e., where there is no component of the deflection in a plane parallel to the surface). The Coriolis 'force' depends on the motion itself. Hence, it affects the direction but not the speed of the air motion, which would involve doing work (i.e., changing the kinetic energy). The Coriolis force always acts at right angles to the direction of the air motion, to the right in the Northern Hemisphere (f positive) and to the left in the Southern Hemisphere (f negative). Absolute values of f vary with latitude as follows:

Latitude	0°	10°	20°	43°	90°
$f(10^{-4}\ s^{-1})$	0	0.25	0.50	1.00	1.458

The earth's rotation also produces a vertical component of rotation about a horizontal axis. This is a maximum at the equator (zero at the poles) and it causes a vertical deflection upward (downward) for horizontal west/east winds. However, this effect is of secondary importance due to the existence of hydrostatic equilibrium.

3 The geostrophic wind

Observations in the *free atmosphere* (above the level affected by surface friction at about 500 to 1000m) show that the wind blows more or less at right angles to the pressure gradient (i.e., parallel to the isobars) with, for the Northern Hemisphere, high pressure on the right and low pressure on the left when viewed downwind. This implies that for steady motion the pressure-gradient force is exactly balanced by the Coriolis deflection acting in the diametrically opposite direction (Figure 6.3A). The wind in this idealized case is called a *geostrophic wind*, the velocity (Vg) of which is given by the following formula:

$$Vg = \frac{1}{2\Omega \sin \phi} \frac{dp}{dn}$$

where dp/dn = the pressure gradient. The velocity is inversely dependent on latitude, such that the same pressure gradient associated with a geostrophic wind speed of 15m s^{-1} at latitude 43° will produce a velocity of only 10m s^{-1} at latitude 90°. Except in low latitudes, where the Coriolis parameter approaches zero, the geostrophic wind

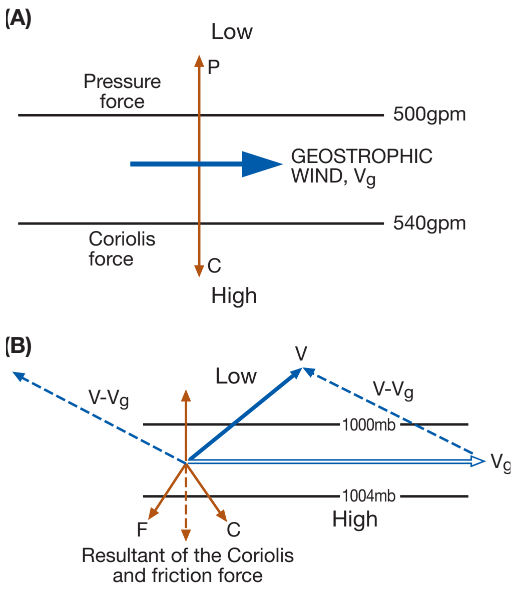

Figure 6.3 A: The geostrophic wind case of balanced motion (Northern Hemisphere) above the friction layer (contour heights are gpm) B: Surface wind **V** represents a balance between the geostrophic wind, $\mathbf{V_g}$, and the resultant of the Coriolis force (**C**) and the friction force (**F**). Note that **F** is not generally directly opposite to the surface wind.

is a close approximation to the observed air motion in the free atmosphere. Since pressure systems are rarely stationary, this fact implies that air motion must continually change towards a new balance. In other words, mutual adjustments of the wind and pressure fields are constantly taking place. The common 'cause-and-effect' argument that a pressure gradient is formed and air begins to move towards low pressure before coming into geostrophic balance is an unfortunate oversimplification of reality.

4 The centripetal acceleration

For a body to follow a curved path there must be an inward acceleration (c) towards the center of rotation. This is expressed by:

$$c = -\frac{mV^2}{r}$$

where m = the moving mass, V = its velocity and r = the radius of curvature. This effect is sometimes regarded for convenience as a centrifugal 'force' operating radially outward (see Note 1). In the case of the earth itself, this is valid. The centrifugal effect due to rotation has in fact resulted in a slight bulging of the earth's mass in low latitudes and a flattening near the poles. The small decrease in apparent gravity towards the equator (see Note 2) reflects the effect of the centrifugal force working against the gravitational attraction directed towards the earth's center. It is therefore only necessary to consider the forces involved in the rotation of the air around a local axis of high or low pressure. Here the curved path of the air (parallel to the isobars) is maintained by an inward-acting, or centripetal, acceleration.

Figure 6.4 shows (for the Northern Hemisphere) that in a low pressure system balanced flow is maintained in a curved path (referred to as the *gradient wind*) by the Coriolis force being weaker than the pressure force. The difference between the two gives the net centripetal acceleration inward. In the high pressure case, the inward acceleration exists because the Coriolis force

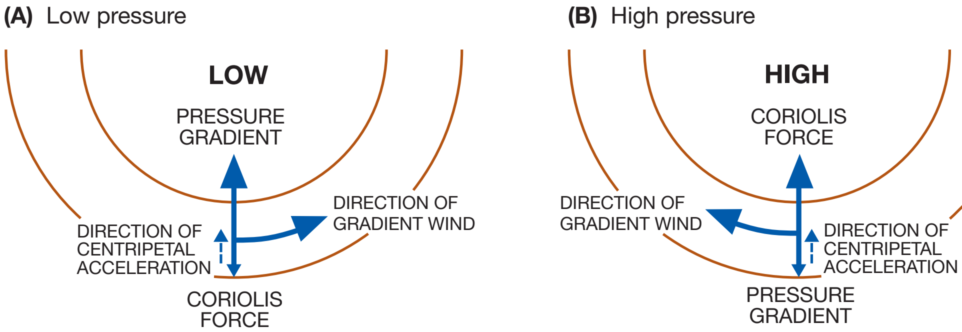

Figure 6.4 The gradient wind case of balanced motion around low pressure (A) and high pressure (B) in the Northern Hemisphere.

exceeds the pressure force. Since the pressure gradients are assumed to be equal, the different contributions of the Coriolis force in each case imply that the wind speed around the low pressure must be lower than the geostrophic value (*subgeostrophic*), whereas in the case of high pressure it is *supergeostrophic*. In reality, this effect is obscured by the fact that the pressure gradient in a high is usually much weaker than in a low. Moreover, the fact that the earth's rotation is cyclonic imposes a limit on the speed of anticyclonic flow. The maximum occurs when the angular velocity is $f/2$ (= V sin ϕ), at which value the absolute rotation of the air (viewed from space) is just cyclonic. Beyond this point anticyclonic flow breaks down ('dynamic instability'). There is no maximum speed in the case of cyclonic rotation.

The magnitude of the centripetal acceleration is generally small, but it becomes important where high-velocity winds are moving in very curved paths (i.e., around an intense low pressure vortex). Two cases are of meteorological significance: first, in intense cyclones near the equator, where the Coriolis force is negligible; and, second, in a narrow vortex such as a tornado. Under these conditions, when the large pressure-gradient force provides the necessary centripetal acceleration for balanced flow parallel to the isobars, the motion is called *cyclostrophic*.

The above arguments assume steady conditions of balanced flow. This simplification is useful, but in reality two factors prevent a continuous state of balance. Latitudinal motion changes the Coriolis parameter, and the movement or changing intensity of a pressure system leads to acceleration or deceleration of the air, causing some degree of cross-isobaric flow. Pressure change itself depends on air displacement through the breakdown of the balanced state. If air movement were purely geostrophic there would be no growth or decay of pressure systems. The acceleration of air at upper levels from a region of cyclonic isobaric curvature (subgeostrophic wind) to one of anticyclonic curvature (supergeostrophic wind) causes a fall of pressure at lower levels in the atmosphere to compensate for the removal of air aloft. The significance of this fact will be discussed in Chapter 9G. The interaction of horizontal and vertical air motions is outlined in B.2, (this chapter).

In cases where the curvature of the flow is tight, as near the eye of a tropical cyclone (see Chapter 11B.2), the centripetal acceleration may balance the pressure gradient force; the resulting wind is termed cyclostrophic.

5 Frictional forces and the planetary boundary layer

The last force that has an important effect on air movement is that due to friction from the earth's surface. Towards the surface (i.e., below about 500m for flat terrain), friction due to form drag over orography begins to reduce the wind velocity below its geostrophic value. This slowing of the wind near the surface modifies the deflective force, which is dependent on velocity, causing it also to decrease. Initially, the frictional force is opposite to the wind velocity, but in a balanced state – when the velocity and therefore the Coriolis deflection decrease (the vector sum of the Coriolis and friction components balances the pressure gradient force (Figure 6.3B). The friction force now acts to the right of the surface wind vector. Thus, at low levels, due to frictional effects, the wind blows obliquely across the isobars in the direction of the pressure-gradient. The angle of obliqueness increases with the growing effect of frictional drag due to the earth's surface averaging about 10–20° at the surface over the sea and 25–35° over land.

In summary, the surface wind (neglecting any curvature effects) represents a balance between the pressure-gradient force and the Coriolis force perpendicular to the air motion, and friction roughly parallel, but opposite, to the air motion. Where the Coriolis force is small, friction may balance the pressure gradient force and the wind (known as antitriptic) flows down the pressure gradient.

Table 6.1 Typical roughness lengths (m) associated with terrain surface characteristics

Terrain surface characteristics	Roughness length (m)
Groups of high buildings	1–10
Temperate forest	0.8
Groups of medium buildings	0.7
Suburbs	0.5
Trees and bushes	0.2
Farmland	0.05–0.1
Grass	0.008
Bare soil	0.005
Snow	0.001
Smooth sand	0.0003
Water	0.0001

Source: After Troen and Petersen (1989).

The layer of frictional influence is known as the *planetary boundary layer* (PBL). Atmospheric profilers (lidar and radar) can routinely measure the temporal variability of PBL structure. Its depth varies over land from a few hundred meters at night, when the air is stable as a result of nocturnal surface cooling, to 1–2km during afternoon convective conditions. Exceptionally, over hot, dry surfaces, convective mixing may extend to 4–5km. Over the oceans it is more consistently near 1km deep and in the tropics especially is often capped by an inversion due to sinking air. The boundary layer is typically either stable or unstable. Yet, for theoretical convenience, it is often treated as being neutrally stable (i.e., the lapse rate is that of the DALR, or the potential temperature is constant with height; see Figure 5.1). For this ideal state, the wind turns clockwise (veers) with increased height above the surface, setting up a wind spiral (Figure 6.5). This spiral profile was first demonstrated in the turning of ocean currents with depth (see Chapter 7D.1a) by V. W. Ekman; both are referred to as *Ekman spirals*. The inflow of air towards the low pressure center generates upward motion at the top of the PBL, known as *Ekman pumping*.

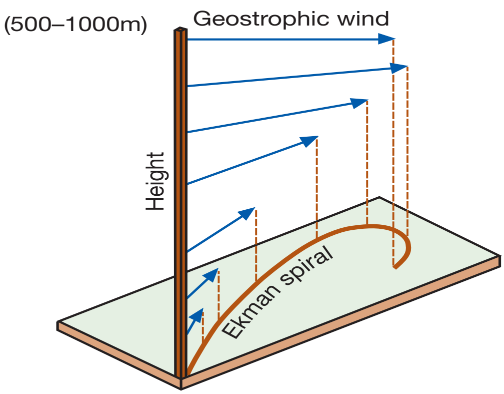

Figure 6.5 The Ekman spiral of wind with height in the Northern Hemisphere. The wind attains the geostrophic velocity at between 500 and 1000m in the middle and higher latitudes as frictional drag effects become negligible. This is a theoretical profile of wind velocity under conditions of mechanical turbulence.

Wind velocity decreases exponentially close to the earth's surface due to frictional effects. These consist of 'form drag' over obstacles (buildings, forests, hills), and the frictional stress exerted by the air at the surface interface. The mechanism of *form drag* involves the creation of locally higher pressure on the windward side of an obstacle and a lateral pressure gradient. Wind stress arises from, first, the molecular resistance of the air to the vertical wind shear (i.e., increased wind speed with height above the surface); such molecular viscosity operates in a laminar sublayer only millimeters thick. Second, turbulent eddies, a few meters to tens of meters across, brake the air motion on a larger scale (eddy viscosity). The aerodynamic roughness of terrain is described by the *roughness length* (z_0), or height at which the wind speed falls to zero based on extrapolation of the neutral wind profile. Table 6.1 lists typical roughness lengths.

Turbulence in the atmosphere is generated by the vertical change in wind velocity, (i.e., a vertical wind shear), and is suppressed by an absence of buoyancy. The dimensionless ratio of buoyant suppression of turbulence to its generation by shear, known as the Richardson number (*Ri*), provides a measure of dynamic stability. Above a critical threshold, turbulence is likely to occur.

B DIVERGENCE, VERTICAL MOTION AND VORTICITY

These three terms are the key to proper understanding of wind and pressure systems on a synoptic and global scale. Mass uplift or descent of air occurs primarily in response to dynamic factors related to horizontal airflow and is only secondarily affected by air-mass stability. Hence the significance of these factors for weather processes.

1 Divergence

Different types of horizontal flow are shown in Figure 6.6A. The first panel shows that air may accelerate (decelerate), leading to velocity divergence (convergence). When streamlines (lines of instantaneous air motion) spread out or squeeze together, this is termed diffluence or confluence, respectively. If the streamline pattern is strengthened by that of the isotachs (lines of equal wind speed), as shown in the third panel of Figure 6.6A, then there may be mass divergence or convergence at a point (Figure 6.6B). In this case, the compressibility of the air causes the density to decrease or increase, respectively. Usually, however, confluence is associated with an increase

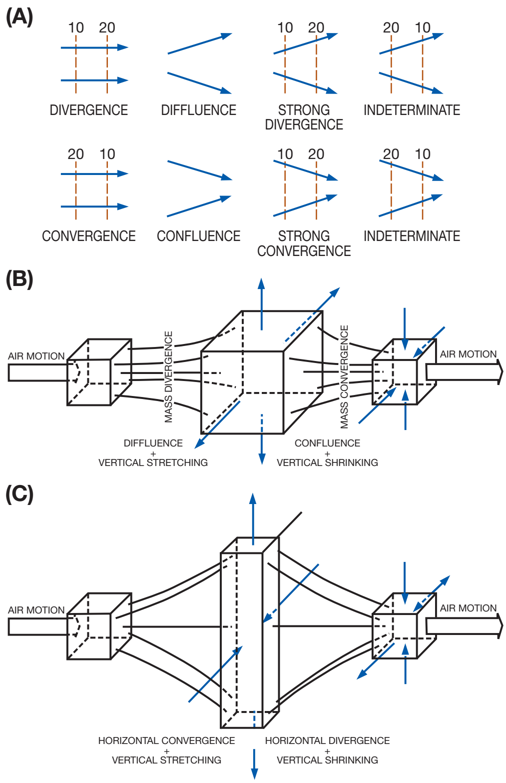

Figure 6.6 Convergence and divergence. A: Plan view of horizontal flow patterns producing divergence and convergence – the broken lines are schematic isopleths of wind speed (isotachs). B: Schematic illustration of local mass divergence and convergence, assuming density changes. C: Typical convergence-stretching and divergence-shrinking relationships in atmospheric flow.

in air velocity and diffluence with a decrease. In the intermediate case, confluence is balanced by an increase in wind velocity and diffluence by a decrease in velocity. Hence, convergence (divergence) may give rise to vertical stretching (shrinking), as illustrated in Figure 6.6C. It is important to note that if all winds were geostrophic, there could be no convergence or divergence and hence no weather!

Convergence or divergence may also occur as a result of frictional effects. Onshore winds undergo convergence at low levels when the air slows down on crossing the coastline owing to the greater friction overland, whereas offshore winds accelerate and become divergent. Frictional differences can also set up coastal convergence (or divergence) if the geostrophic wind is parallel to the coastline with, for the Northern Hemisphere, land to the right (or left) of the air current, viewed downwind.

2 Vertical motion

Horizontal inflow or outflow near the surface has to be compensated by vertical motion, as illustrated in Figure 6.7, if the low or high pressure systems are to persist and there is to be no continuous density increase or decrease. Air rises above a low pressure cell and subsides over high pressure, with compensating divergence and convergence, respectively, in the upper troposphere. In the middle troposphere, there must clearly be some level at which horizontal divergence or convergence is effectively zero; the mean 'level of non-divergence' is generally at about 600mb. Large-scale vertical motion is extremely slow compared with convective up- and downdrafts in cumulus clouds, for example. Typical rates in large depressions and anticyclones are of the order of ±5–10cm s^{-1}, whereas updrafts in cumulus may exceed 10m s^{-1}.

3 Vorticity

Vorticity implies the rotation, or angular velocity, of small (imaginary) parcels in any fluid. The air within a low pressure system may be regarded as comprising an infinite number of small air parcels, each rotating cyclonically around an axis vertical to the earth's surface (Figure 6.8). Vorticity has three elements – magnitude (defined as *twice* the angular velocity, Ω) (see Note 3), direction (the

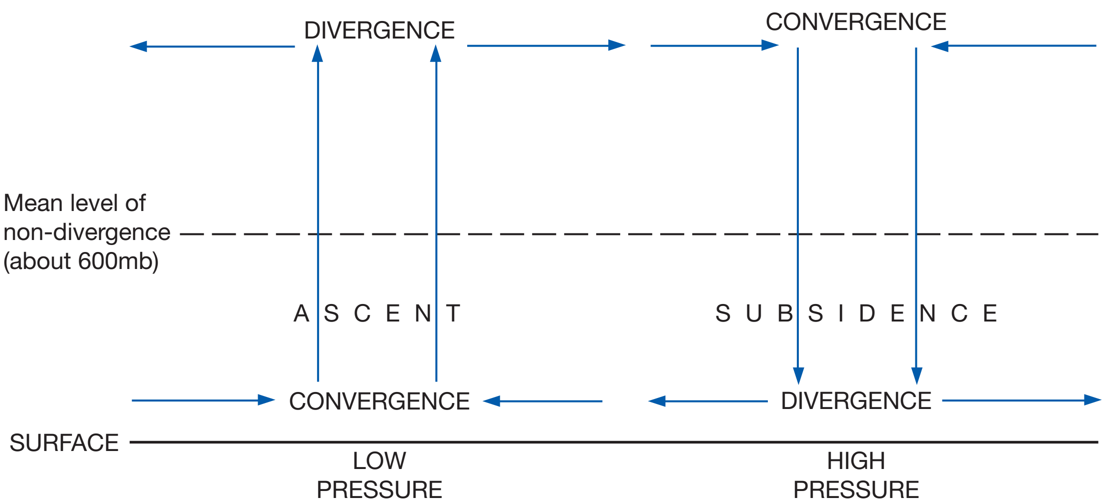

Figure 6.7 Cross-section of the patterns of vertical motion associated with (mass) divergence and convergence in the troposphere, illustrating mass continuity.

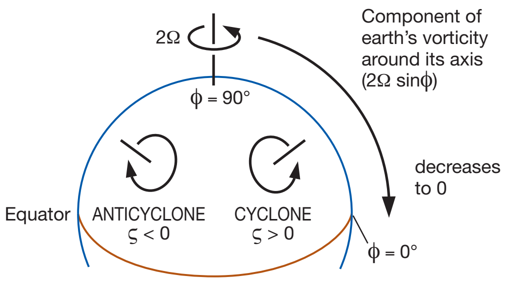

Figure 6.8 Sketch of the relative vertical vorticity (ζ) about a cyclone and an anticyclone in the Northern Hemisphere. The component of the earth's vorticity around its axis of rotation (or the Coriolis parameter, f), is equal to twice the angular velocity (Ω) times the sine of the latitude (f). At the pole $f = 2\Omega$, diminishing to 0 at the equator. Cyclonic vorticity is in the same sense as the earth's rotation about its own axis, viewed from above, in the Northern Hemisphere: this cyclonic vorticity is defined as positive ($\zeta > 0$).

horizontal or vertical axis around which the rotation occurs) and the sense of rotation. Rotation in the same sense as the earth's rotation – cyclonic in the Northern Hemisphere – is defined as positive. Cyclonic vorticity may result from cyclonic curvature of the streamlines, from cyclonic shear (stronger winds on the right side of the current, viewed downwind in the Northern Hemisphere), or a combination of the two (Figure 6.9). Lateral shear (see Figure 6.9B) results from changes in isobar spacing. Anticyclonic vorticity occurs with the corresponding anticyclonic situation. The component of vorticity around an axis vertical to the earth's surface is referred to as the vertical vorticity. This is generally the most important, but near the ground surface frictional shear causes vorticity around an axis parallel to the surface and normal to the wind direction.

Vorticity is related not only to air motion around a cyclone or anticyclone (*relative vorticity*), but also to the location of that system on the rotating earth. The vertical component of *absolute vorticity* consists of the relative vorticity (ζ) and the latitudinal value of the Coriolis parameter, $f = 2\Omega \sin \phi$ (see Chapter 6A). At the equator, the

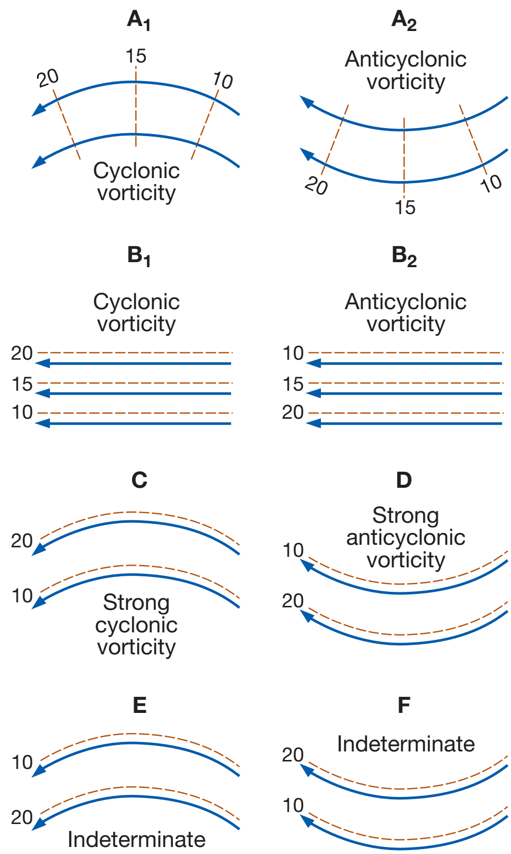

Figure 6.9 Streamline models illustrating in plan view the flow patterns with cyclonic and anti-cyclonic vorticity in the Northern Hemisphere. In C and D, the effects of curvature (a_1 and a_2) and lateral shear (b_1 and b_2) are additive, whereas in E and F they more or less cancel out. Dashed lines are schematic isopleths of wind speed.

Source: After Riehl *et al*. (1954).

local vertical is at right angles to the earth's axis, so $f = 0$, but at the North Pole cyclonic relative vorticity and the earth's rotation act in the same sense (see Figure 6.8) and $f = 2\Omega$.

C LOCAL WINDS

For a weather observer, local controls of air movement may present more problems than the effects of the major planetary forces just discussed. Diurnal tendencies are superimposed upon both the large- and the small-scale patterns of wind velocity. These are particularly noticeable in the case of local winds. Under normal conditions, wind velocities tend to be least about dawn when there is little vertical thermal mixing and the lower air is less affected by the velocity of the air aloft (see Chapter 7A). Conversely, velocities of some local winds are greatest around 13:00–14:00 hours, when the air is most subject to terrestrial heating and vertical motion, thereby enabling coupling to the upper air movement. Air always moves more freely away from the surface, because it is not subject to the retarding effects of friction and obstruction.

Table 6.2 gives a summary classification of local winds, each of which is now discussed in detail.

1 Mountain and valley winds

Terrain features give rise to their own special meteorological conditions. On warm, sunny days, the heated air in a valley is laterally constricted compared with that over an equivalent area of lowland, and so tends to expand vertically. The volume ratio of lowland/valley air is typically about 2 or 3:1 and this difference in heating sets up a density and pressure differential, which causes air to flow from the lowland up the axis of the valley. This valley wind (Figure 6.10) is generally light and requires a weak regional pressure gradient in order to develop. This flow along the main valley develops more or less simultaneously with *anabatic* (upslope) winds, which result from greater heating of the valley sides compared with the valley floor. These slope winds rise above the ridge tops and feed an upper return current along the line of the valley to compensate for the valley wind. This feature may be obscured, however, by the regional airflow. Speeds reach a maximum around 14:00 hours.

At night, there is a reverse process as denser cold air at higher elevations drains into depressions and valleys; this is known as a *katabatic* wind. If the air drains downslope into an open valley, a 'mountain wind' develops more or less simultaneously along the axis of the valley. This flows towards the plain, where it replaces warmer, less dense air. The maximum velocity occurs just

Table 6.2 Classification of local winds

Name	Characteristics	Forcing
Anabatic	Daytime upslope warm flow	Horizontal density gradient towards the slope
Katabatic	Night-time downslope cold flow	Gravity and horizontal density gradient away from the slope
Mountain wind	Night-time cold down-valley flow	Mountains-to-plains density gradient
Valley wind	Daytime warm up-valley flow	Plains-to-mountains density gradient
Anti-mountain wind	Above the mountain wind in the opposite direction	Compensation current
Anti-valley wind	Above the valley wind in the opposite direction	Compensation current
Sea breeze	Day-time flow from the seas to the land	Density gradient from cool sea to heated land
Land breeze	Night-time flow from land to sea	Density gradient from cool land to warmer sea
Foehn (Chinook)	Down lee slope with increasing T and lower RH	Blocked flow on windward side; or flow crossing mountains with cloud/precipitation on windward slope
Bora	Down lee slope with air colder than that it replaces	Blocked flow of cold air upwind
Barrier wind	Low-level flow parallel to the mountains, directed poleward	Blocking reduces the flow speed normal to the barrier reducing the Coriolis force

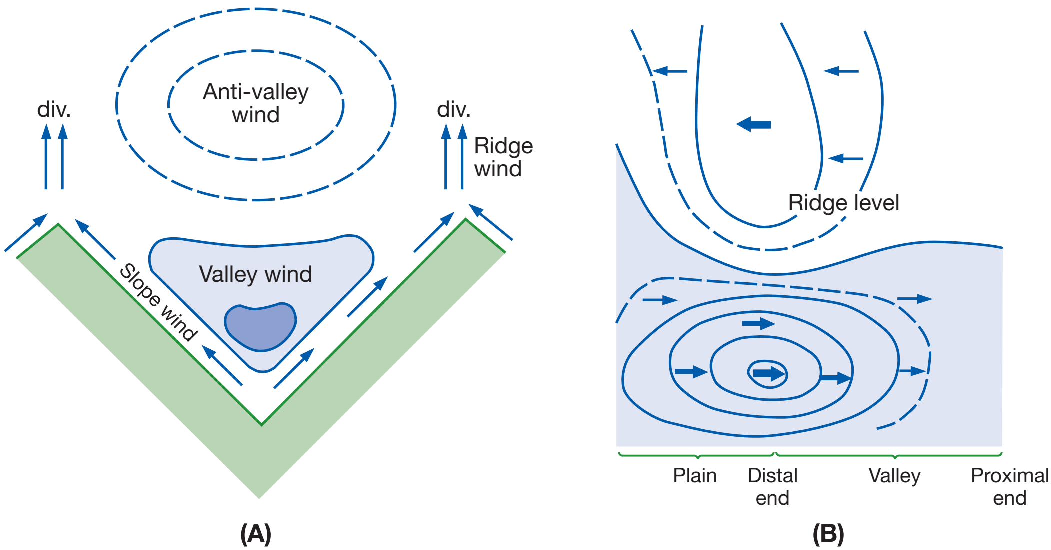

Figure 6.10 Valley winds in an ideal V-shaped valley. A: Section across the valley. The valley wind and anti-valley wind are directed at right angles to the plane of the paper. The arrows show the slope and ridge wind in the plane of the paper, the latter diverging (div.) into the anti-valley wind system. B: Section running along the center of the valley and out on to the adjacent plain, illustrating the valley wind (below) and the anti-valley wind (above).

Source: After Buettner and Thyer (1965).

before sunrise at the time of maximum diurnal cooling. As with the valley wind, an upper return current, in this case up-valley, also overlies the mountain wind.

Katabatic drainage is usually cited as the cause of frost pockets in hilly and mountainous areas. It is argued that greater radiational cooling on the slopes, especially if they are snow-covered, leads to a gravity flow of cold, dense air into the valley bottoms. Observations in California and elsewhere, however, suggest that the valley air remains colder than the slope air from the onset of nocturnal cooling, so that the air moving downslope slides over the denser air in the valley bottom. Moderate drainage winds will also act to raise the valley temperatures through turbulent mixing. Cold air pockets in valley bottoms and hollows probably result from the cessation of turbulent heat transfer to the surface in sheltered locations rather than by cold air drainage, which is often not present.

2 Land and sea breezes

Another thermally induced wind regime is the land and sea breezes (see Figure 6.11). The vertical expansion of the air column that occurs during daytime heating over the more rapidly heated land (see Chapter 3B.5) tilts the isobaric surfaces downward at the coast, causing onshore winds at the surface and a compensating offshore movement aloft. Typical land–sea pressure differences are of the order of 2mb. At night, the air over the sea is warmer and the situation is reversed, although this reversal is also the effect of downslope winds blowing off the land. Figure 6.12 shows that sea breezes can have a decisive effect on temperature and humidity on the coast of California. A basic offshore gradient flow is perturbed during the day by a westerly sea breeze. Initially, the temperature difference between the sea and the coastal mountains of central California sets up a shallow sea breeze, which by midday is

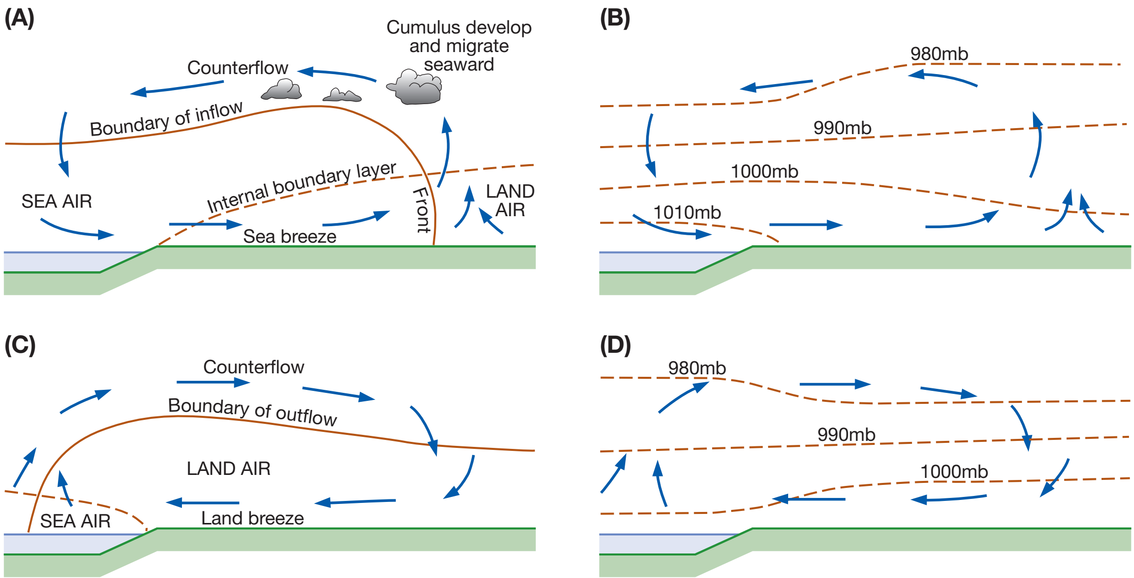

Figure 6.11 Diurnal land and sea breezes. A and B: sea breeze circulation and pressure distribution in the early afternoon during anticyclonic weather. C and D: land breeze circulation and pressure distribution at night during anticyclonic weather.

Source: A and C after Oke (1978).

300m deep. In the early afternoon, a deeper regional-scale circulation between the ocean and the hot interior valleys generates a 1km-deep onshore flow that persists until two to four hours after sunset. Both the shallow and the deeper breeze have maximum speeds of 6m s^{-1}. A shallow evening land breeze develops by 19:00 PST but is indistinguishable from the gradient offshore flow.

The advancing cool sea air may form a distinct line (or *front*; see Chapter 8D) marked by cumulus cloud development, behind which there is a distinct wind velocity maximum. This often develops in summer, for example, along the Gulf coast of Texas. On a smaller scale, such features are observed in Britain, particularly along the south and east coasts. The sea breeze has a depth of about 1km, although it thins towards the advancing edge. It may penetrate 50km or more inland by 21:00 hours. Typical wind speeds in such sea breezes are 4–7m s^{-1}, although these may be greatly increased where a well-marked low-level temperature inversion produces a 'Venturi effect' by constricting and accelerating the flow. The much shallower land breezes are usually weaker, about 2m s^{-1}. Counter-currents aloft are generally weak and may be obscured by the regional airflow, but studies on the Oregon coast suggest that under certain conditions this upper return flow may be very closely related to the lower sea breeze conditions, even to the extent of mirroring the surges in the latter. In mid-latitudes the Coriolis deflection causes turning of a well-developed onshore sea breeze (clockwise in the Northern Hemisphere) so that eventually it may blow more or less parallel to the shore. Analogous 'lake breeze' systems develop adjacent to large inland water bodies such as the Great Lakes and even the Great Salt Lake in Utah.

Small-scale circulations can be generated by local differences in albedo and thermal conductivity. Salt flats (playas) in the western deserts of the United States and in Australia, for example, cause on off-playa breeze by day and an on-playa flow at night due to differential heating. The salt

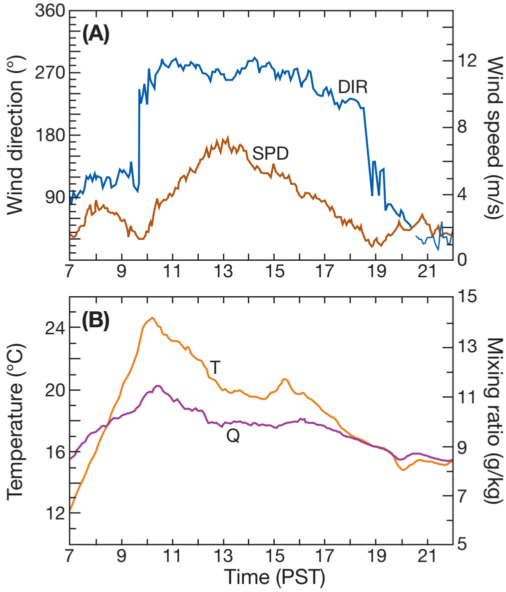

Figure 6.12 The effects of a westerly sea breeze on the California coast on 22 September 1987 on temperature and humidity. Above: Wind direction (DIR) and speed (SPD); below: air temperature (T) and humidity mixing ratio (Q) on a 27m mast near Castroville, Monterey Bay, California. The gradient flow observed in the morning and evening was easterly.

Source: Banta (1995, p. 3621, Fig. 8).

flat has a high albedo and the moist substrate results in a high thermal conductivity relative to the surrounding sandy terrain. The flows are about 100m deep at night and up to 250m by day.

3 Winds due to topographic barriers

Mountain ranges strongly influence airflow crossing them. On the upwind side of mountains perpendicular to the airflow, blocking may occur when the airflow is stable and unable to cross the barrier. As the flow approaches the barrier it slows down, thus reducing the Coriolis force. Imbalance with the pressure gradient force then causes the air to turn poleward towards the lower pressure on the left side of the flow. This sets up a low-level *barrier wind* that may feature a low-level (850mb) jet of 20m s^{-1}. Such winds are common upstream of the Sierra Nevada, California.

The displacement of air upward over an obstacle may trigger instability if the air is conditionally unstable and buoyant (see Chapter 5B), whereas stable air returns to its original level in the lee of a barrier as the gravitational effect counteracts the initial displacement. This descent often forms the first of a series of *lee waves* (or *standing waves*) downwind, as shown in Figure 6.13. The wave form remains more or less stationary relative to the barrier, with the air moving quite rapidly through it. Below the crest of the waves, there may be circular air motion in a vertical plane, which is termed a *rotor*. The formation of such features is of vital interest to pilots. The presence of lee waves is often marked by the development of lenticular clouds, and on occasion a rotor causes reversal of the surface wind direction in the lee of high mountains.

Winds on mountain summits are usually strong, at least in middle and higher latitudes. Average speeds on summits in the Colorado Rocky Mountains in winter months are around 12–15m s^{-1}, for example, and on Mount Washington, New Hampshire, an extreme value of 103m s^{-1} has been recorded. Peak speeds in excess of 40–50m s^{-1} are typical in both of these areas in winter. Airflow over a mountain range causes the air below the tropopause to be compressed and thus accelerated particularly at and near the crest line (the Venturi effect), but friction with the ground also retards the flow, compared with free air at the same level. The net result is predominantly one of retardation, but the outcome depends on the topography, wind direction and stability.

Over low hills, the boundary layer is displaced upward and acceleration occurs just above the summit. Figure 6.14 shows instantaneous airflow conditions across Askervein Hill (relief *c.* 120m) on the island of South Uist in the Scottish Hebrides, where the wind speed at a height of 10m above the ridge crest approaches 80 percent more than the undisturbed upstream velocity. In

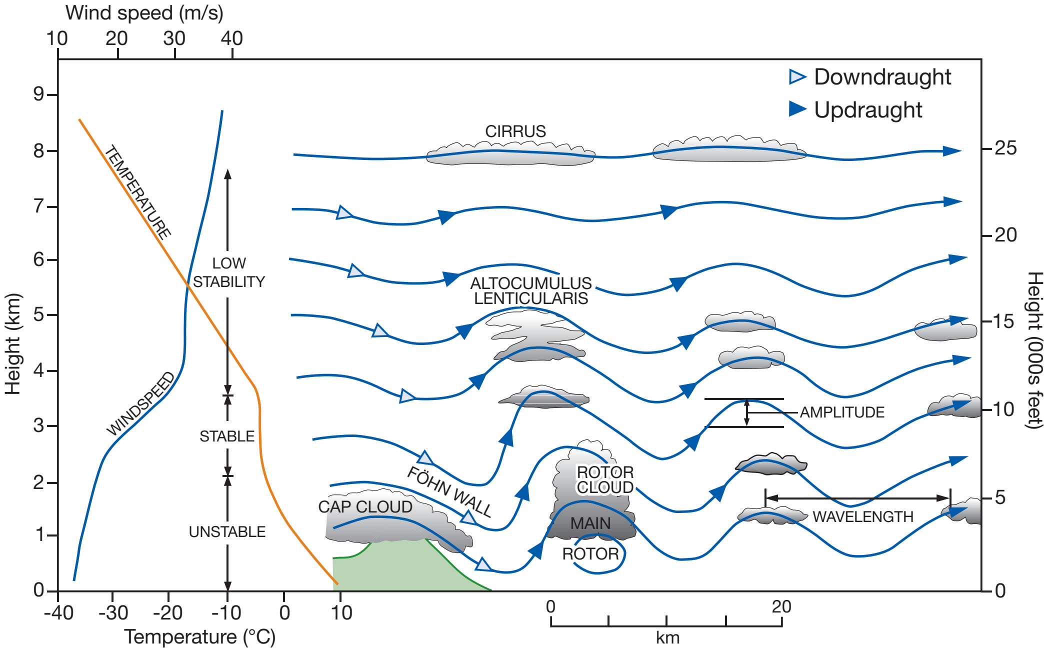

Figure 6.13 Lee waves and rotors are produced by airflow across a long mountain range. The first wave crest usually forms less than one wavelength downwind of the ridge. There is a strong surface wind down the lee slope. Wave characteristics are determined by the wind speed and temperature relationships shown schematically on the left of the diagram. The existence of an upper stable layer is particularly important.

Source: After Ernst (1976).

contrast, there was a 20 percent decrease on the initial run-up to the hill and a 40 percent decrease on the lee side, probably due to horizontal divergence. Knowledge of such local factors is critical for siting wind energy systems.

A wind of local importance near mountain areas is the *föhn*, or *chinook*. It is a strong, gusty, dry and warm wind that develops on the lee side of a mountain range when stable air is forced to flow across the barrier by the regional pressure gradient; the air descending on the lee slope warms adiabatically. Sometimes, there is a loss of moisture by precipitation on the windward side of the mountains (Figure 6.15). The air, having cooled at the saturated adiabatic lapse rate above the condensation level, subsequently warms at the greater dry adiabatic lapse rate as it descends on the lee side. This also reduces both the relative and the absolute humidity. Other investigations show that in many instances there is no loss of moisture over the mountains. In such cases, the föhn effect is the result of the blocking of air to windward of the mountains by a summit-level temperature inversion. This forces air from higher levels to descend and warm adiabatically. Southerly föhn winds are common along the northern flanks of the Alps and the mountains of the Caucasus and Central Asia in winter and spring, when the accompanying rapid temperature rise may help to trigger avalanches on the snow-covered slopes. At Tashkent in Central Asia, where the mean winter temperature is around freezing point, temperatures may rise to more than 21°C during a föhn. In the same way, the chinook is a significant feature at the eastern foot of the New Zealand Alps, the Andes in Argentina, and the Rocky

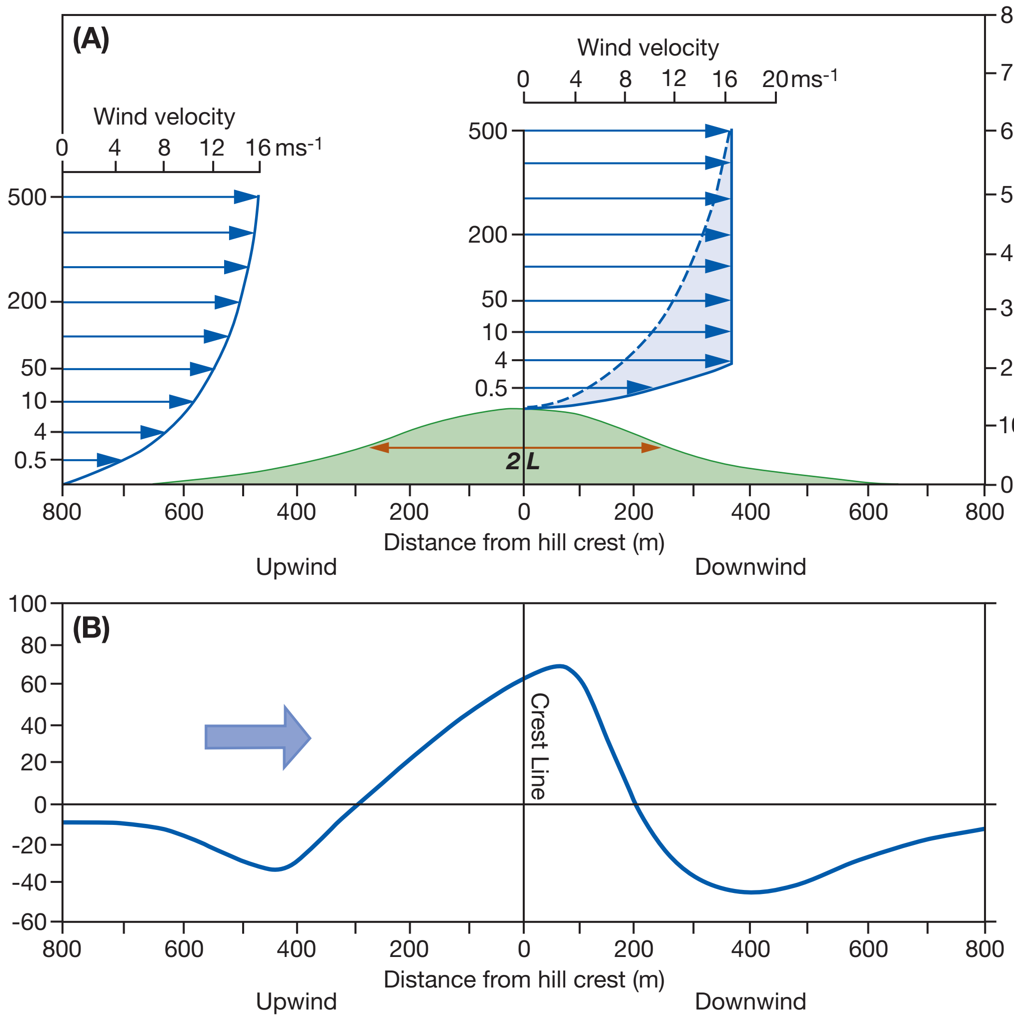

Figure 6.14 Airflow over Askervein Hill, South Uist, off the west coast of Scotland. A: Vertical airflow profiles (not true to scale) measured simultaneously 800m upwind of the crest line and at the crest line. L is the *characteristic length* of the obstruction (i.e. one-half the hill width at mid-elevation, here 500m) and is also the height above ground level to which the flow is increased by the topographic obstruction (shaded). The maximum speed-up of the airflow due to vertical convergence over the crest is to about 16.5m s^{-1} at a height of 4m. B: the relative speed-up (%) of airflow upwind and downwind of the crest line measured 14m above ground level.

Source: After Taylor, Teunissen and Salmon *et al*. From Troen and Petersen (1989).

Mountains. At Pincher Creek, Alberta, a temperature rise of 21°C occurred in four minutes with the onset of a chinook on 6 January 1966. In California, the Santa Ana is a cold season easterly wind that blows from the deserts east of the Sierra Nevada to the coast of southern California. It has an average frequency of 20 events per year and average duration of 1.5 days. It is notable for the dry air, which greatly increases the risk of chaparral fires. Less spectacular effects are also

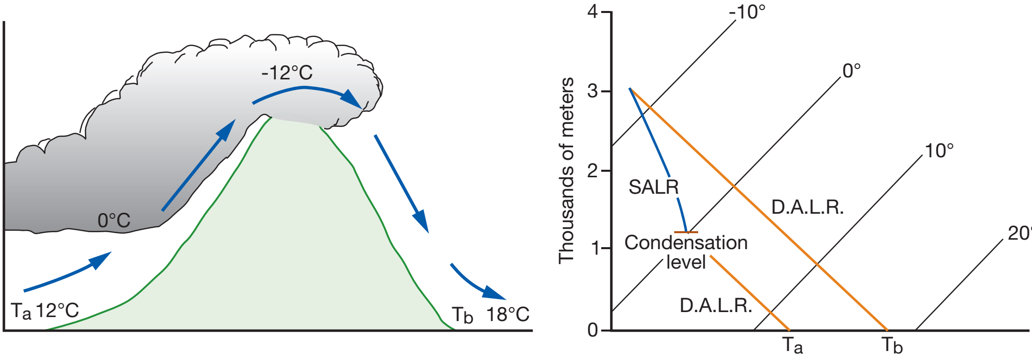

Figure 6.15 The föhn effect when an air parcel is forced to cross a mountain range 3km high. T_a refers to the temperature at the windward foot of the range and T_b to that at the leeward foot.

noticeable in the lee of the Welsh mountains, the Pennines and the Grampians in Great Britain, where the importance of föhn winds lies mainly in the dispersal of cloud by the subsiding dry air. This is an important component of so-called 'rain shadow' effects.

In some parts of the world, winds descending on the lee slope of a mountain range are colder than the air they displace (despite adiabatic warming through descent). The type example of such 'fall-winds' is the *bora* of the northern Adriatic, where cold northeasterly flows cross the Dinaric Alps, although similar winds occur on the northern Black Sea coast, in northern Scandinavia, in Novaya Zemlya and in Japan. These winds occur when cold continental air masses are forced across a mountain range by the pressure gradient and, despite adiabatic warming, displace warmer air. They are therefore primarily a winter phenomenon.

On the eastern slope of the Rocky Mountains in Colorado (and in similar continental locations), winds of either bora or chinook type can occur depending on the initial airflow characteristics. Locally, at the foot of the mountains, such winds may attain hurricane force, with gusts exceeding 45m s^{-1} (100mph). Down-slope storms of this type have caused millions of dollars of property damage in Boulder, Colorado, and the immediate vicinity. These wind storms develop when a stable layer close to the mountain-crest level prevents air to windward from crossing over the mountains. Extreme amplification of a lee wave (see Figure 6.13) drags air from above the summit level (4000m) down to the plains (1700m) over a short distance, leading to high velocities. However, the flow is not simply 'down-slope'; winds may affect the mountain slopes but not the foot of the slope, or vice versa, depending on the location of the lee wave trough. High winds are caused by the horizontal acceleration of air towards this local pressure minimum.

Air motion is described by its horizontal and vertical components; the latter are much smaller than the horizontal velocities. Horizontal motions compensate for vertical imbalances between gravitational acceleration and the vertical pressure gradient.

The horizontal pressure gradient, the earth's rotational effect (Coriolis force), and the curvature of the isobars (centripetal acceleration) determine horizontal wind velocity. All three factors are accounted for in the gradient wind equation, but this can be approximated in large-scale flow by the geostrophic wind relationship. Below 1500m, the wind speed and direction are affected by surface friction.

Air ascends (descends) in association with surface convergence (divergence) of air. Air motion is also subject to relative vertical vorticity as a result of curvature of the streamlines and/or lateral shear; this, together with the earth's rotational effect, makes up the absolute vertical vorticity.

Local winds occur as a result of diurnally varying thermal differences setting up local pressure gradients (mountain–valley winds and land–sea breezes) or due to the effect of a topographic barrier on airflow crossing it (examples are the lee-side föhn and bora winds).

- Compare the wind direction and speed reported at a station near you with the geostrophic wind velocity determined from the MSL pressure map for the same time (data sources are listed in Appendix 4).
- Why would there be no 'weather' if the winds were strictly geostrophic?
- What are the causes of mass divergence (convergence) and what roles do they play in weather processes?
- In what situations do local wind conditions differ markedly from those expected for a given large-scale pressure gradient?

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