Chapter 5: The atmosphere in motion: winds, weather, and the general circulation of the atmosphere

5.1 The forces that drive atmospheric motion

If we look at the atmosphere at the molecular scale (10⁻¹⁰ m), we see individual molecules travelling at the speed of sound (~400 m s⁻¹), until they bounce off another air molecule or a solid surface. On larger spatial scales, we observe the atmosphere behaving like a fluid, with air currents that appear to be organized on all spatial scales, from less than 1 m to global. Velocities for these motions range from less than 0.01 m s⁻¹ in still air, up to hundreds of m s⁻¹ in tornadoes.

In this chapter we examine the forces that drive atmospheric motions, that organize the motions and produce weather. We also study the largest scale atmospheric motions, the general circulation of the global atmosphere, and we begin to think about climate, long-term statistical averages of weather that define habitats for plants and animals.

The most basic driver of atmospheric motion is differential heating, which makes air at one location warmer than at another. Differential heating can be associated with sunlight heating different features of the landscape, or with evaporation or condensation of water leading to storage of energy as latent heat vaporization. The following factors, discussed in detail in this chapter, regulate atmospheric winds:

1. **Heating** of the earth's surface or the atmosphere, by sunlight or by infrared (heat) radiation, causes air temperatures to vary from place to place. Through the Perfect Gas Law, these temperature variations lead to corresponding differences in atmospheric pressure (*pressure gradients*) and density (*density gradients*). Movement of mass from/into a vertical column of air also gives rise to pressure and density gradients. Temperature changes may give rise to buoyancy if the lapse rate becomes steeper than the adiabatic lapse rate, leading to convection.

2. **Pressure gradients** cause air to begin to move from the region of high pressure to the region of low pressure, just as air or water may be forced through a hose by applying pressure to one end.

3. The **Coriolis force** turns the flow of air, to the right in the Northern Hemisphere and the left in the Southern Hemisphere. The Coriolis force is a phenomenon, due to the rotation of the earth, which makes moving objects appear to follow curved trajectories on the earth's surface.

4. Other energy sources that can modify the flow of air include release of latent heat, absorption or emission of radiant heat by air molecules or by particles suspended in the air, obstacles such as mountains, and friction against the earth's surface.
5.2 Pressure gradients and a simple circulation

*Pressure Force*

A pressure difference between two points generates a *pressure force* that causes the air to flow from the higher-pressure point to the lower-pressure point.

\[ \text{A } \rightarrow \text{ B} \]

If \( P_A > P_B \), there will be a pressure force \( F \) directed from A to B, and according to Newton's 2\textsuperscript{nd} Law, air parcels will be accelerated from A towards B.

Note: There is an important difference between the effect of changing pressure in the horizontal, versus the vertical, direction. The weight of the overlying atmosphere is balanced by changes in pressure in the vertical (barometric law). Vertical motions are caused by *deviations* from the balance of forces of the barometric law, especially due to buoyancy. In the horizontal direction, there is no force of gravity and even a slight pressure gradient can produce wind. (A *pressure gradient* is defined as the change in pressure \( P \) over a given distance \( S \), \( \Delta P/\Delta S \)). This example emphasizes the need to think about both the magnitude and the direction of a force (see Fig. 5.1).

![Figure 5.1 Net force on an object.](image)

*Sea breeze circulation*

We illustrate the effect of differential heating in producing pressure gradients and winds by studying the *sea breeze* phenomenon, shown in Figure 5.2, and discussed in connection with a demo (differentially heated tank) in Chapter 4. Consider a coastline on a nice sunny day during the summer. The sun heats the land surface to a higher temperature than the ocean, due largely to the fact that sunlight is mostly absorbed right at the land surface, but in the ocean light is absorbed over a layer meters (up to 100 meters) deep. The higher temperature of the air over the land means that this air has lower density than adjacent parcels over the sea, and it becomes buoyant (see Chapter 4). The buoyant air over the land tends to rise.

Warming of the air over the land has a second, equally important effect. As the air column over
land warms up, it expands to have a larger scale height \( H (=kT/mg) \) than over the sea. Thus the pressure falls off more slowly with altitude over land and a gradient of pressure develops at higher altitude. The direction of this pressure gradient is opposite to the density difference at the surface, with the higher pressure over the land (see Figure 5.2). This pressure gradient causes mass to be transferred from the land to the sea at altitude, adding mass to the air column over the sea and raising the pressure at the sea surface. The higher pressure over the sea at the surface forces air from sea to land. Pressure gradients are set up over land and sea as shown in Figures 5.2 and 5.3.

**Fig. 5.2 Illustration of the sea breeze**, showing the circulation and the relative pressures in the horizontal direction, near the ground and aloft (see also Fig. 5.3). The land heats up during the daytime, but the sea does not. Due to the higher temperature, the atmosphere over the land has lower density and a larger scale height \( H \) than over the sea. The lower density makes air over the land buoyant relative to air over the sea, and it rises. The larger scale height makes the pressure aloft higher over the land than over the sea, causing mass to be transferred from land to the sea at altitude. The associated addition of mass to the air column over the sea raises the pressure at the sea surface, setting up the distribution of high and low pressure in Figure 5.3 and giving rise to the circulation shown.

**Figure 5.3. Distribution of pressure with altitude in the sea breeze.** The atmosphere expands as it is heated over the land, increasing the scale height \( H \), reducing the rate of pressure decline with altitude. At altitude, the pressure is higher over land than over the adjacent sea, which causes mass to be transferred to the air column over the sea. Surface pressure over the ocean is therefore increased, giving rise to the distribution of pressure shown in the figure.

Figures 5.2 and 5.3 illustrate the winds and pressure gradients set up by this sequence of events, the sea breeze. Dry convection over land takes mass and energy from the surface to higher altitudes, setting up a horizontal pressure gradient aloft that pushes mass from the land to the sea, and the resultant pressure gradient at the surface pushes air from sea to land. A circulation is set up, in which air parcels follow a loop course, rising over land, moving out to sea at altitude, descending over the sea, and moving from sea to land at the surface. The sea breeze is limited in
extent: ~ 1-10 km in the horizontal, and about 1 km deep. It is one of the simplest examples of convection due to buoyancy, with resulting pressure gradients and circulation, generated by differential heating in the atmosphere. It is an analogue for important features of the global circulation that we discuss in a later chapter.

5.3 Coriolis Force

Nature of the Coriolis force.

When you set an object in motion on a rotating platform (for example, throwing the ball in the demonstration with two students tossing a ball on a rotating platform), it appears to follow a curved path. It behaves as if a force were pulling it, with the direction and strength of the force related to the direction and speed of rotation and of the object. Actually, the object is travelling in a straight line in space, and it is we, the observers rotating with the platform, who are travelling along a curved trajectory.

The Coriolis force is a "fictitious" force introduced by physicists to help describe this situation. It is a force that modifies the motion of objects in the frame of reference of a rotating sphere (which is the frame of reference for us, citizens of the rotating Earth!).

Let's say that you are on Earth at latitude $\lambda$ (Greek lambda). The rotation of the Earth around its polar axis makes you travel every day a distance corresponding to the perimeter of the circle of radius $R \cos(\lambda)$ (Fig. 5.4, $R = 6370$ km). You thus travel at a speed

$$v = \frac{2\pi R \cos(\lambda)}{t}$$

(Eq. 5.1)

where $t = 1$ day (86400 seconds). The latitude of Boston is 42°; plugging in numbers, you will
find that you are traveling at a constant speed \( v = 1250 \text{ km/h (800 mph!)} \). You are oblivious to this rapid motion because all the objects in your frame of reference (the people you see, the buildings, the trees, etc.) are traveling at the same speed. However, from the perspective of an observer fixed in space, you are indeed traveling very fast round and round the Earth. Note that \( V \) decreases with increasing latitude, and that therefore points at different latitudes on Earth are moving on circles of different sizes, at different speeds.

*Trajectories of projectiles on the earth*

To understand the Coriolis Force, let's consider an observer shooting a ball at a target. Assume first that this observer is at the North Pole and shoots at a target fixed at lower latitude. Looking down on the pole, we see a view as in Figure 5.5 (top panel).

![Figure 5.5. Illustration of the Coriolis force. (Left panel). An observer sitting on the axis of rotation (North Pole) launches a projectile at the target. The curved arrow indicates the direction of rotation of the earth. (Right panel) The projectile follows a straight-line trajectory, when viewed by an observer in space, directed towards the original position of the target. However, observers and target are rotating together with the earth, and the target moves to a new position as the projectile travels from launch to target. Since observers on earth are not conscious of the fact that they and the target are rotating with the planet; they see the projectile initially heading for the target, then veering to the right. The Coriolis force is a fictitious force introduced to the equations of motion for objects on a rotating planet, sufficient to account for the apparent pull to the right in the Northern hemisphere or to the left in the southern hemisphere.](image)

It will take a finite time \( \Delta t \) for the ball to travel from the observer to the target. During that time the target will have moved (due to the rotation of the Earth), causing the ball to miss the target. The observer at the North Pole does not perceive the target as having moved, because everything in the observer's frame of reference is moving in the same way. However, the observer can clearly see that the shot has missed. From this perspective, the ball has been deflected to the right. Such a deflection implies a force (the Coriolis force) being exerted to the right of the direction of motion. This force is fictitious in that a second observer fixed in space would not see
any such deflection (see Figure 5.5).

Consider now the case of an observer at latitude near the equator shooting a ball at a target at higher latitude, as illustrated in Figure 5.6.

Figure 5.6. Illustration of the apparent deflection of a projectile launched from low latitude (equator) to high latitude in the Northern Hemisphere. The observer launches the projectile with northward velocity, but because the observer is travelling with the rotation of the earth, the projectile also has a (very large) velocity from west to east. The target is closer to the axis of rotation of the earth, and it therefore is not travelling a rapidly from west to east. The projectile moves ahead of the target as it moves north, and it therefore appears to the observer to be deflected to right. A projectile launched towards a target in the Southern Hemisphere would appear to be deflected to the left.

The translational (longitudinal) velocity \( V \) of the rotating Earth is higher at \( \lambda_1 \) than \( \lambda_2 \). If the momentum of the ball is conserved, then as the ball travels northward it is moving longitudinally faster than its surrounding environment, so that it is deflected to the right of the target.

We may examine this process from the point of view of angular momentum. As the ball travels from low latitude towards the pole, it decreases its distance from the axis of rotation of the earth. Since the ball must conserve its angular momentum \( \text{its angular velocity} \) (number of revolutions around the axis of the earth per unit time) increases. We are familiar with this principle in everyday life, for example when skaters or ballet dancers increase the rate of a spin by pulling in their arms.

**Exercise:** Draw diagrams similar to 5.5 and 5.6 for the Southern Hemisphere.

The angular momentum \( L \) for mass \( m \) rotating at angular velocity \( \Omega(= 2 \pi / \tau \text{ where } \tau = \text{period of rotation}) \) is given by

\[
L = mr^2 \Omega.
\]  

(Eq. 5.2)

Here \( r \) is the distance from the axis of rotation. The requirement that \( L \) be conserved implies that, if \( r \) changes, \( \Omega \) must change so as to counteract the change in \( r \), i.e. \( \Omega = L/(mr^2) \). For example, if \( r \) were to decrease by factor 2, \( \Omega \) would increase by factor 4 so that \( L \) would stay unchanged.

The Coriolis force also applies to an object launched at a target located at the same latitude. Consider a ball thrown from west to east in the Northern Hemisphere. The ball is travelling
longitudinally faster than its environment and hence experiences a larger *centrifugal force*, perpendicular to the axis of rotation of the Earth, than stationary objects. Because the earth's surface is curved, this centrifugal force can be decomposed into a vertical component pointing upwards from the surface and a horizontal component pointing to the right of the direction of motion, as shown in Figure 5.7.

![Diagram showing centrifugal force](image)

The horizontal component deflects the ball to the right. The vertical component deflects the ball upwards, but this effect is negligible when compared to the other large vertical forces applied to the ball (pressure force, gravity).

**Exercise:** show that the ball would still be deflected to the right if it were thrown from east to west rather than from west to east.

We thus find in all cases that the Coriolis force is exerted *perpendicular to the direction of motion*, to the **RIGHT** in the Northern Hemisphere and to the **LEFT** in the Southern Hemisphere.

![Diagram showing deflection](image)

**Figure 5.8.** When a projectile is launched towards the east in the Northern Hemisphere, it appears to be deflected to the right. The object has a greater angular velocity than stationary objects on the earth's surface, and therefore a greater centrifugal force. The component of that extra force oriented along the earth's surface pulls the object towards the equator (to the right). Had the object been launched towards the west, it would have curved towards the pole (also deflected to the right).

Using Newton's Laws, we can show that the magnitude of the Coriolis force is given by
Here $m$ is the mass of the moving object, $\Omega = 2\pi/\tau = 7.29 \times 10^{-5}$ radian s$^{-1}$ is the *angular velocity* of the Earth, $\tau$ is the period of rotation (1 day), and $v$ is the speed of the moving object relative to the earth's surface (not the speed of the Earth!). The *Coriolis acceleration* is obtained by dividing the force by the mass (since $F = ma$):

$$\text{Coriolis acceleration} = F/m = 2\Omega v \sin(\lambda).$$  
(Eq. 5.4)

Note that the Coriolis force increases as $\lambda$ (latitude) increases and is zero at the equator (convince yourself that the Coriolis force must indeed be zero at the equator).

The Coriolis force is relatively weak, and objects travelling short distances are not noticeably deflected (see problem xx). The Coriolis force can be very important however for large-scale motions, including for example, the trajectory of artillery shells (problem xz). From the above expression for the Coriolis acceleration one can calculate the displacement $\Delta y$ incurred when throwing an object with speed $v$ at a target at distance $\Delta x$. The calculation requires a little bit of calculus and so we give the result here:

$$\Delta y = \Omega (\Delta x)^2 / v \sin(\lambda)$$

At Boston ($\lambda = 42^\circ$N), we find that a snowball traveling 10 m at 20 km/h is displaced by $\Delta y = 1$ mm (negligible), but a missile traveling 1000 km at 2000 km/h is shifted 100 km (important!).

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*Demonstrations for Coriolis force and angular momentum*

- **Rotating-platform demo.** Students
  1. **shoot baskets**
  2. **toss the ball vertically into the air**

while seated on a rotating platform. The ball appeared to the players to curve to the right if the platform rotated counter-clockwise, to the left if it rotated clockwise. Observers in the room saw the ball going in a straight line, but the players' aim appeared to be off.

This demo illustrates the *Coriolis force*, an apparent acceleration that makes moving objects follow curved trajectories when viewed from a rotating platform (like the earth!).

- **Angular momentum vector property demo.** Students experience a force when they try
to change the axis of rotation of a bicycle wheel. The direction of the force depends on which way they turned the axis of the wheel, and on the direction that the wheel was spinning. The angular momentum of the wheel is a vector quantity, and when the student is standing on a platform free to rotate, conservation of angular momentum makes the student spin when trying to re-orient the wheel.

This demo illustrates angular momentum, shows the importance of direction of rotation, and illustrates by direct experience the vector properties.

- **Angular momentum conservation.** The "dumbbells" experiment shows a student on the rotating platform spinning faster when s/he pulls in the weights.

This demo shows that objects increase their angular velocity (number of revolutions per second) if they are moved closer to the axis of rotation, and vice versa. Thus a wind blowing from north to south in the Northern Hemisphere slows its angular velocity and appears to observers on the earth to be deflected to the right. A wind blowing from south to north increases its angular velocity, also appear to be to be deflected to the right.

*The relative motions of a moving parcel and the earth's surface are illustrated using a globe, ball-on-a-string, etc.*

### 5.4 The atmosphere in motion: geostrophic flow, atmospheric weather systems

**Geostrophy and circulation around highs and lows**

Air flows in the atmosphere in response to forces acting on it. We saw in the previous lecture how the pressure force resulting from differences in heating across a coastline produces the familiar sea breeze. We also saw how the Coriolis force could modify the direction of atmospheric flow over large scales. The pressure force and the Coriolis force are the most important and ubiquitous forces acting in the horizontal direction in the atmosphere. As we will see, they act together to determine the strength and direction of the winds, with implications for the formation of clouds and rain.

*Fig. 5.9a. Motion of an air parcel in the Northern Hemisphere, starting at rest and subjected to a north/south pressure gradient as in Fig. 5.9b.*
Consider an example of a pressure difference between two regions of the atmosphere, created for example by a difference in heating, with higher pressure to the south. Figure 5.9 illustrates a pressure difference, showing lines of constant pressure called isobars (contours of constant pressure). Isobars are frequently drawn on weather maps, and we will shortly see why.

Consider now a parcel of air initially at rest at point A\(_1\). This air parcel experiences a pressure force pointing towards the region of low pressure. It does not experience a Coriolis force because it is at rest (the Coriolis force applied to an object is proportional to its speed, see Eq. 5.3). The pressure force acting on the air parcel orients itself towards the direction of sharpest decrease in pressure and is therefore perpendicular to the isobars. Due to this force, the air parcel is accelerated towards the region of low pressure and starts to move at increasing speed. As it acquires speed, it experiences a Coriolis force acting perpendicular to the direction of motion and towards the right (in the Northern Hemisphere). This force will cause the air parcel to deviate to the right from the direction of the pressure force (Fig. 5.9a, point A\(_2\) in Figure 5.9b). As the air parcel continues to accelerate, the Coriolis force increasingly pulls the air parcel to the right.

If we keep the pressure gradients fixed, an air parcel starting from rest as depicted in Fig. 5.9 will continue to be deflected to the right until it is travelling against the pressure gradient; then it will slow down due to the opposing pressure force. It will undergo an oscillatory motion, and it is not actually transported along the pressure gradient. It does however drift to the right (from West to East, in the example of Figure 5.9). In the real atmosphere, the oscillations will be damped out, because air will re-arrange itself in the atmosphere to construct a new pressure gradient for which the associated parcel motion and Coriolis force exactly cancel. The only the net motion of the air is then along (parallel to) the isobars (point A\(_3\)). In this state, with the Coriolis force equal in magnitude and in the opposite direction to the pressure force, air cannot flow from high to low pressure. The cancellation of pressure and Coriolis forces leaves no net force acting on the air, and air movement is steady and parallel to the isobars (zero acceleration).

The steady wind resulting from the balance of the pressure force and the Coriolis force is called the geostrophic flow. The corresponding wind speed depends on the strength of the pressure force (that is, how closely spaced the isobars are). A larger pressure force requires a larger...
Coriolis force and hence a higher wind speed in order to achieve the balance of forces necessary for geostrophic flow (see problem 5 below).

As we have seen from our example, the atmosphere tends naturally towards the equilibrium represented by the geostrophic flow. Observations show that winds indeed generally approximate geostrophic flow, thus the reason for putting isobars on weather maps. From the orientation of the isobars we can infer the direction of the wind (parallel the isobars), and from strength of the gradient (closeness of isobars to one another) we know the strength of the wind (the more closely spaced the isobars the stronger the wind).

The orientation of geostrophic flow allows us to determine the direction of winds that air flows around a center of high pressure (clockwise around a "High") or a center of low pressure (counterclockwise around a "Low") in the northern hemisphere. The directions of flow are, of course, reversed in the Southern Hemisphere (see Figure 5.10).

Example: Velocity of winds in geostrophic balance

Problem: Find the speed and direction of the geostrophic wind for typical atmospheric conditions, as in the following: latitude 30°N, density \( \rho = 0.7 \) kg m\(^{-3}\), at a pressure of 500 mb (approximately 6 km altitude). The pressure gradient is 4 mb over a distance of 100 km.

Solution: We approach this problem in two steps. First, we find the pressure force on an air parcel. Second, we use the formula giving the relationship between Coriolis force and velocity to find the velocity just sufficient to cancel out the pressure force.

Step 1. Assume that the drop in pressure is linear with distance. The change in pressure across a distance of 1 m is obtained by dividing the total change in pressure (4 mb) by the distance (10\(^5\) m), 4\times10\(^{-5}\) mb m\(^{-1}\). This defines the magnitude of the pressure gradient \( \Delta P/\Delta s \), where \( \Delta P \) denotes the change in pressure observed over distance \( \Delta s \). The conversion factor is 10\(^5\) N m\(^{-2}\) = 1000 mb, or 10\(^2\) N m\(^{-2}\) mb\(^{-1}\), so \( \Delta P/\Delta s = 4\times10^{-3} \) N m\(^{-3}\). Thus there is a force of 0.004 Newtons on each cubic meter of air.

Step 2. The Coriolis force on the same cubic meter is given by 2 \( \Omega \rho v_g \sin(\lambda) \) (Eq. 5.3), where \( v_g \) is the wind speed and \( \lambda \) is latitude. If we are to obtain geostrophic balance, the two forces must be equal in magnitude and opposite in direction. Hence we wish to find the velocity \( v_g \) giving the Coriolis force equal to the pressure force,

\[
\Delta P/\Delta s = 2 \Omega \rho v_g \sin(\lambda).
\]

Solving for \( v_g \), we find

\[
v_g = \frac{\Delta P / \Delta s}{2 \Omega \rho \sin(\lambda)}.
\]

With \( \Omega = 7.3\times10^{-5} \) s\(^{-1}\), \( \rho = 0.7 \) kg m\(^{-3}\), \( \sin(\lambda) = 0.5 \), \( v_g = 4\times10^{-3}/(2\times7.3\times10^{-5}\times0.7\times0.5) = 78.2 \) m s\(^{-1}\).

The speed of the geostrophic wind is 78 m s\(^{-1}\), oriented along the isobar, with the high pressure to the right of the direction of motion (high pressure would be to the left in the Southern Hemisphere).
Relationship between isobars and geopotentials on weather maps.

Weather maps in newspapers or on TV often show contours of geopotential height rather than contours of pressure. The geopotential height is the altitude at which a given pressure is found in the atmosphere. For example, we may see a weather map showing contours of geopotential height for 850 mb, i.e. the altitude at each point on the map at which this pressure is found. The maps showing contours of geopotential height look very similar to maps showing isobars, with highs and lows in approximately the same places. The figures and text in this box show the relationships between isobars and geopotential heights, and help explain why the geostrophic wind flows parallel to contours of geopotential height much as winds flow parallel to isobars.

![Diagram showing the relationship between pressure and geopotential height](image)

Figure showing the relationship between pressure and geopotential height. The pressures and corresponding geopotential heights for two points s1 and s2 on a weather map (the map is shown on the next figure) are shown schematically. The figure shows the pressure versus height for each point. Note the higher pressure at s1 than at s2. Symbols are keyed to small section of the map shown below. The geopotential heights for pressure P1 are Z1 (■) and Z2 (□) respectively, whereas the pressures at height Z1 are P1 (■) and P2 (Δ) respectively. Note that the pressure being higher at S1 than S2 (P1>P2), requires that likewise the geopotential is higher at S1 than at S2 (Z1>Z2).
Schematic maps of isobars at a constant altitude $Z_1$ (lines in the left panel) and of corresponding geopotentials for pressure $P_1$ (lines in the right panel). The contour labeled $P_1$ and passing through point $s_1$ represents a pressure higher than $P_2$ at $s_2$; likewise the contour labeled $Z_1$ and passing through point $s_1$ represents a geopotential height higher than $Z_2$ at $s_2$.

**Figure 5.10.** Circulation of air around regions of high and low pressures in the Northern Hemisphere. *Upper panel:* A region of high pressure produces a pressure force directed away from the high. Air starting to move in response to this force is deflected to the right (in the Northern Hemisphere), giving a clockwise circulation pattern.

*Lower panel:* A region of low pressure produces a pressure force directed from the outside towards the low. Air starting to move in response to this force is also deflected to the right, rotating counter-clockwise.

Directions of rotation of the wind about high or low centers are reversed in the Southern Hemisphere, as explained earlier in this chapter.
**Divergence and convergence of air in atmospheric highs and lows**

A geostrophic wind has a peculiar property: since it blows parallel to the isobars, no mass is transferred from regions of higher pressure to regions of lower pressure. We know however that highs and lows persist only a few days, and that particular weather conditions are associated with highs (fair, dry) and lows (cloudy, precipitation). In fact, winds are only approximately geostrophic and air is generally transported, albeit rather slowly, from regions of high to regions of low pressure. Near the surface of the Earth, a third force needs to be considered that plays an important role in a variety of weather phenomena, including modification of the geostrophic flow: the friction force. Air loses momentum as it travels horizontally near the surface due to collision with trees, buildings, and other obstacles. This loss of momentum implies a force exercised in a direction opposite to the direction of motion and which we call the *friction force*.

**Fig. 5.11a.** The effect of friction around a high pressure region is to slow the wind relative to its geostrophic velocity. This causes the pressure force to slightly exceed the Coriolis force. The three forces add together as shown in the figure. Air parcels gradually drift from higher to lower pressure, in the case shown here, from the center of a high pressure region outward. An analogous flow (inward) occurs in a low-pressure region. See Figure 5.11b.

In order to balance the triangle of three forces (pressure, Coriolis, friction) for air flowing horizontally near the surface, the flow around a region of high pressure must be deflected away from the High (divergence) while the flow around a region of low pressure is deflected towards the Low (convergence). This result is the same in the North or South Hemisphere, even though the direction of circulation is opposite..
Figure 5.11b. Air converges near the surface in low pressure centers, due to the modification of geostrophic flow under the influence of friction (see Fig. 5.11a). Air diverges from high pressure centers. At altitude, the flows are reversed: divergence and convergence are associated with lows and highs respectively, closing the circulation through analogous processes noted in the sea breeze example (Section 5.2).

Air must sink from above in order to replenish the outflow of mass caused by air moving away from a high-pressure center at the surface. As this air sinks, it's compressed and heats up, and relative humidity decreases; weather conditions are usually sunny and dry. Conversely, rising motions balance convergence of air at the surface in a low-pressure region, and adiabatic cooling of the rising air often causes formation of clouds and rain. Hence high pressure is generally associated with fair weather, and low pressure with wet weather, in both hemispheres.

Maps and pictures of Hurricane Josephine (October, 1996)

The following images and maps chart the progress of Hurricane Josephine in October 1996, intended to familiarize students with presentation of weather information and to visualize the concepts of atmospheric circulation discussed above. The storm started northward 7 October 1996 from the Gulf of Mexico, moved across Florida and up the East Coast, reaching New England on 8 September 1996. We show weather maps illustrating the circulation around this very strong low-pressure center, with corresponding satellite pictures showing clouds and rain.

These weather maps show geopotential height, defined as the altitude at which a given pressure is located, instead of isobars at a given altitude (see discussion box above). Maps of geopotential height have very similar interpretation to maps showing isobars. In a region of relatively low pressure, we have to go to lower altitude to find a given pressure (e.g. 850 mb, near-surface values used in the maps below), and correspondingly for high pressure and higher altitude.
Figure 5.12a. Pressure and winds for 07 October 1996. Colors show surface pressure (mean sea level, difference between observed barometric pressure and 1014 mb). The arrows show the direction and magnitude of the wind 10 m above ground level. See if you can identify the following features on this map: (1) counterclockwise circulation ("cyclonic circulation") around the hurricane (deep low); (2) convergence towards the center of the low, instead of wind blowing directly parallel to the isobars; (3) location of the strongest winds in the areas with strong gradients of pressure. Data source: NCEP Daily Analysis, Diagnostic Mean Sea Level Pressure and Mean Wind at 10m above ground [http://ingrid.ldeo.columbia.edu].

Figure 5.12b. Infrared photograph of Hurricane Josephine on 07 October 1996 (Fig. 5.12a) from the NOAA GOES-8 weather satellite. Color added to the cloud features enhance the image, with colder clouds (higher in the atmosphere) shown as brighter. The brightest clouds are very cold and deep, and are producing intense rainfall in the storm center and in a band of strong rainfall across the Gulf of Mexico, extending to the Pacific.
Figure 5.12c. Progress of the storm. One day later, the hurricane has moved to the Carolinas. The prominent features of the cyclonic circulation are evident (see caption to Figure 5.12a). Note the strongest winds in the northeast sector.

Figure 5.12d. Infrared picture (without color enhancement) about 24 hours after the time shown in Figure 5.12c. The storm lies just off the coast of New England. It has weakened, as have the rain bands trailing towards the southwest.
5.5 The atmosphere in motion: the general circulation of the atmosphere

Figure 5.13 Mean observed fields of surface pressures and wind directions around the globe, January and July. Cyclonic (counter-clockwise) circulations around low-pressure regions are apparent, reverse around high-pressure. The trade winds (easterly) prevail in the tropics. Circulation in summer differs from winter, especially in the Northern Hemisphere. Note the winter highs over continents and lows over oceans, reversed in summer, and the Indian monsoon in summer. Data source: NCEP reanalyzed fields (for surface pressure) and A. Oort climatology (for winds) [http://ingrid.ldeo.columbia.edu].
Observed features of the general circulation of the atmosphere.

Atmospheric weather systems come and go, winds shift, temperatures rise and fall from hour to hour, day to day. The weather is notoriously variable and difficult to predict. Nevertheless, if we observe the atmosphere over long periods of time, persistent and very important patterns emerge. The long-term average patterns of wind and temperature, as well as the statistics summarizing many episodes that deviate from these averages (persistence of droughts, flooding events, etc) represent the climate. Particular weather events usually are unimportant for man or the environment (except for very extreme events), but changes in climate can be very important, with long-term implications for plants and animals, including humans.

The progress of the seasons is one of the most obvious of climate phenomena. Our experience readily distinguishes weather events from the progression of the seasons. For example, in New England in 1998, several days in December were warmer than in August. On average, of course, we expect early winter conditions in December (e.g. significant snowfall), but not in August. The plants and animals of the region know to expect this also, and millions of years of evolution have insured that trees are in full leaf in August but bare in December. An anomalous day of weather is insignificant, and such anomalies abound. However long-term shifts in climate (temperature, precipitation) or changes in the frequency or severity of extreme events, can have major impacts on natural ecosystems and on human beings, leading to replacement of the current assemblages of plants and animals by new ones.

Figure 5.13 shows averages of highs and lows and associated winds over the globe, calculated from actual data over a decade. We note a number of characteristic features:

- At many places near the equator, southeasterly and northeasterly winds converge at the intertropical convergence zone ("ITCZ"). This feature is usually apparent even on singly images of the globe, a ribbon of atmosphere, only a few hundred km wide, with persistent clouds and rain. For example, the ITCZ is shown in the global satellite images in Chapter 1, and it shows up again as a series of storms across the bottom of the satellite images in Figure 5.12. The latitude of the ITCZ on average follows the latitude of maximum solar heating, moving to about 12N in July and 3S in January. The asymmetry (12 vs. 3 degrees from the equator) reflects the asymmetry in the distribution of landmasses between north and south.

- North and south of the ITCZ easterly trade winds, extending to about 20-30 degrees latitude. (Winds blowing from east to west are called easterly winds.) Sailors since Columbus used the trade winds to cross the Atlantic from Europe to the Americas. Trade winds are present also over much of the Pacific.

- At 20-30 degrees north and south are regions of prevailing high pressure, with well-defined centers of high pressure (subtropical anticyclones) over the oceans and over eastern Asia in winter. The divergent flow in high pressure regions (see Fig. 5.11 above) warms the air as it descends (air parcels try to follow the adiabatic lapse rate) leading to
low relative humidity, and dry conditions. Divergent, anti-cyclonic flow is apparent around the high-pressure centers in Figure 5.13 (clockwise in the Northern Hemisphere, counter-clockwise in the Southern Hemisphere). Thus high pressure is associated with dry conditions, and the major deserts of the world are in the belt 20-30° latitude.

- At higher latitudes (mid-latitudes) the winds shift to a prevailing westerly direction. This flow is stronger in the Southern Hemisphere (e.g., the "roaring forties") than in the Northern Hemisphere.

- There is remarkable symmetry in the large-scale features on both sides of the equator, especially if one accounts for seasonal changes. Comparison between the January and July maps indicates a number of seasonal differences, similar, but not identical in the two hemispheres. A notable example is the monsoon circulation over India, present in July (follows solar heating!) but not in January; a monsoon is the low pressure, rising air over continents in summer, due to solar heating of the land. The great scarp of the Himalayas amplifies the tendency for air to be warmed over the continents in summer, so there is no analogue to the Indian monsoon in the Southern Hemisphere.

The Hadley Circulation, and modification of Hadley's concepts by the Coriolis effect

The first model for explaining the general circulation of the atmosphere was discussed by Hadley in 1735. He envisioned the circulation as a "global sea breeze", driven by the contrast between the hot tropics and the cold poles, as shown in Fig. 5.14.
This model accounts for a remarkable number of features of the General Circulation known to sailors in the 18th century, including the presence of the ITCZ near the equator and the seasonal movement of the ITC from the southern tropics in January to the northern tropics in July. However, a crucial problem in the Hadley model is that it does not account properly for easterlies and westerlies, i.e. for the effect of the Coriolis force; in fact, Hadley advanced incorrect arguments to account for these well-known aspects of the circulation.

The core concept in Hadley's physical model is correct. The General Circulation of the atmosphere is indeed driven by heating in the tropics, and cooling at high latitudes. However, Coriolis force deflects the high-altitude branch of the Hadley circulation as it moves poleward. Simple calculations show that (either Exercise or problem for this), by the time the flow reaches 30° latitude, wind speeds become excessive, approaching the speed of sound. Instead of simply continuing northward, the winds turn westerly and a geostrophic balance is approximately set up. The flow is roughly parallel to lines of latitude, and does not go further poleward. Thus the circulation envisioned by Hadley does not extend to latitudes higher than 30°. At this point the air has nowhere to go but down. Thus the Hadley circulation can only extend from the equator to about 30 degrees; at 30 degrees the air is pushed down, producing the subtropical high-pressure belts seen in the observations (Figures 5.1 and 5.15).
Figure 5.15a. (Left) Deflection of the upper branch of the Hadley circulation by the Coriolis force.

Figure 5.15b. (below) Schematic picture of the General Circulation of the atmosphere, showing the Hadley circulation between ±30° latitude and the effect of the Coriolis force on the return flow at the surface, giving rise to the easterly trade winds in the tropics. The westerlies at middle latitudes are depicted as arising from the high-latitude branches of the subtropical high pressure regions. The mid-latitude or polar jet streams are shown occupying the regions of strongest temperature and pressure gradients at the mid-latitude/polar latitude boundaries.

Figure 5.15b shows a highly schematic picture of the General Circulation of the atmosphere (compare to 5.13). The Hadley circulation is restricted to ±30° latitude. Note that the return flow near the surface is deflected to the right, becoming easterly. This circulation is consistent with the motion ("anti-cyclonic, clockwise in the Northern Hemisphere, counterclockwise in the Southern Hemisphere) expected around the subtropical high pressure zones. Hence the Hadley circulation explains the easterly trade winds when we take account of the Coriolis force. The observed fact that the trade winds are restricted to low latitudes reflects the limitation imposed by the Coriolis force, preventing the Hadley circulation from extending much beyond 30°.
latitude. The Westerly winds at mid-latitudes can be viewed as spinning off from the poleward branch of this anticyclonic flow, also responding to the Coriolis force.

Poleward transport of air (likewise heat and moisture) from the tropics to latitudes higher than 30° is inhibited by the strength of the Coriolis force. The pressure gradient pushing air polewards tends to be balanced by the Coriolis force at middle and high latitudes, resulting in westerly flow. If the earth rotated more slowly and the Coriolis force were correspondingly weaker, we might find the Hadley circulation extending to the poles, and the Earth's climate would be very different, with the polar regions much warmer than today. Indeed, Venus rotates very slowly and the Hadley circulation appears to extend to the pole, helping to make the temperature difference very small between equator and pole.

In fact at mid-latitudes the prevailing flow is westerly at all altitudes. In order to push the air (and hence heat) poleward, it is essential for the air to slow down and to lose the angular momentum it carries from lower latitudes. This is accomplished in the Northern Hemisphere by the presence of large land masses and mountains, causing friction. Atmospheric motions at middle latitudes feature strong disturbances in the mean flow, which we experience as changes in wind direction and propagating high and low pressure zones (storms). The variations occur with typical time scales of 2-5 days (called the synoptic time scale), and the underlying swirling motions bear some resemblance to the swirls that appear in flowing water, and are therefore called atmospheric eddies. Movement of heat by eddy motions is more efficient in the Northern Hemisphere than in the Southern Hemisphere, where the surface is mainly relatively smooth ocean. This is why one observes very strong, persistent westerly winds in southern mid-latitudes. The Antarctic atmosphere is particularly cold because strong winds that are approximately geostrophic do not transport mass (or heat) to higher latitudes; most of the heat and mass is transported by unsteady eddy motions.

The polar jet stream forms where the Coriolis force inhibits the atmosphere from transporting heat to the poles, especially in winter. A strong temperature gradient (the "polar front") develops between 30 N (Florida) and 45 N (Minneapolis), especially over land, due to rapid cooling at high latitudes by radiation of heat to space. A strong pressure difference is caused at high altitude by this temperature gradient (Figure 5.16), giving rise to a strong westerly flow.

The meandering of the jet stream brings alternating warm and cold weather to middle latitudes in winter, e.g. in Boston. This process produces the typical weather pattern we experience in winter: alternating cold, clear, dry periods (anti-cyclonic flow, winds from the northwest) and warmer stormy periods (winds from the South or East), and these storms and cold outbreaks represent the synoptic eddies mentioned above. They are helps to transporting heat polewards (warm air streaming north, cold air flowing south), taking over where the Hadley circulation leaves off.
Other features of the general circulation

- The monsoons of India, North America and Eurasia are the seasonal reversal of low and high-pressure areas over the continents, caused by the fact that the continental surface responds more strongly to heating in summer and cooling in winter. Low pressure is found over the continent in summer for the same reason that there is low pressure over land in the sea breeze: the land is warmer than the surrounding ocean. However the spatial scale is much larger and the Coriolis effect leads to the circulation shown in the pressure map for July. The reverse is true in winter.

- Near the equator during winter, a principal feature of the circulation in the tropics is a set of 3 low-pressure (warmer) regions: the Western Pacific, Equatorial Africa, and Amazon Basin. The latter two are fixed geographically by the continents, where solar heating produces warmer air relative to the oceans, and thus upwelling and rain. The warmest continental regions with the largest land areas and latent heat fluxes produce atmospheric upwelling (similar to the sea breeze, on continental scale), and cooler areas with less release of latent heat have descending air that closes this circulation. The warmest/wettest region of all is in the Pacific, usually associated with the large islands of Indonesia, Malaysia, and New Guinea. This 3-zone circulation at equatorial latitudes produces a pattern of alternating lows (heaviest rain) and highs (dry). Most of the downwelling that closes the circulation occurs over the oceans. In some cases however this downwelling occurs over land, causing the semi-arid equatorial landscapes in Eastern Brazil and East Africa.

- Sometimes the warm region of rainfall and atmospheric upwelling in the Pacific migrates far to the east, following a pool of very warm ocean water. This shift has global climate effects, and is called the El Niño phenomenon (also known as the El Niño-Southern Oscillation or ENSO). When this happens, the distributions of highs and lows is changed over the whole globe--California gets rain, Indonesia, Australia, and parts of Brazil get droughts, and sometimes the northeastern US gets a warm, snow-free winter.
Scientists have made significant progress in forecasting the occurrence of El Niño by observing the location of the warmest water in the Pacific, giving 3-4 months notice of the coming climatic anomaly. The year 1997 presented one of the strongest El Niño events on record, and the effects of fires in unusually dry Indonesia and floods in unusually wet California made world news and caused vast damage to human society.

The El Niño-Southern Oscillation has occurred every few years in recent decades, but during some epochs may have not occurred at all. Climates were likely quite different at those times than at present.

A movie and satellite images are available on (http://www.courses.fas.harvard.edu/~scia30/) that provide visualization of the global circulation. These images show several important aspects of the global atmospheric motions on particular days, and therefore not all of the features are apparent. A remarkable number are clearly visible however.

Main points of Chapter 5.

Sections 5.1-5.3 Pressure gradients, sea breeze, and Coriolis force.

- Differential heating produces changes in pressure through the barometric law, and pressure forces set the atmosphere into motion.

- The sea breeze develops due to heating of the land during the day, while the sea hardly changes temperature. A simple sequence of events produces a circulation with air rising over land, blowing from land to sea above the surface, descending over the sea, and blowing from sea to land (the "sea breeze") at the surface.

- The Coriolis force is an apparent acceleration that causes moving objects on a rotating platform (e.g., the earth) to follow trajectories that appear curved to an observer on the rotating platform. The Coriolis force on the Earth's surface is given by the formula

\[ F = 2m\Omega v \sin(\lambda) \]

where \(\lambda\) is latitude, \(\Omega\) is the angular velocity, and \(v\) is the velocity of the moving object as viewed by an observer rotating with the earth (one of us!). Air deflected to the right in the Northern Hemisphere and to the left in the Southern Hemisphere

- The deflection of an object due to the Coriolis force larger for longer distances, i.e. in the atmosphere, the curvature is greater for big objects like major storms:

\[ \Delta y = \Omega (\Delta x)^2 / v \sin(\lambda) \]

Section 5.4 Geostrophy and circulation around highs and lows

- Due to the Coriolis force, winds blow counterclockwise around a low pressure region
(cyclonic circulation) in the Northern Hemisphere and clockwise around a high pressure region (anti-cyclonic circulation). The terms "cyclonic" and anti-cyclonic" apply to circulation about lows and highs in the Southern Hemisphere, but the sense of rotation is reversed.

• Above the surface (away from the influence of friction and convection), the pressure gradient force is usually approximately balanced by the Coriolis force, a condition called **geostrophy**. Geostrophic winds blow *parallel to lines of constant pressure (isobars at a fixed altitude) and to lines of constant geopotential height (at fixed pressure)*. Winds are not generally *exactly* geostrophic, due to changing conditions, momentum exchange (friction) with the surface, etc., and thus are not exactly parallel to isobars.

• Near the surface (or near the influence of convection), the pressure gradient force, the Coriolis force, and friction are approximately balanced, leading to inflow of air near the surface in a low pressure region (convergence) or outflow in a high pressure region (divergence). Convergence and divergence force vertical motion, up and down respectively, giving precipitation (lows) or clear weather (highs).

**Section 5.5 The General Circulation of the atmosphere: atmospheric motions on large spatial scales and long time scales**

• The **Hadley Circulation** represents the effects of heating at the equator and cooling at the poles, leading to global upwelling at equatorial latitudes (the **Intertropical Convergence Zone, (ITCZ)**) and downwelling (arid conditions) in the subtropics.

• The Hadley circulation is limited to ±30 degrees of latitude due to the influence of the Coriolis force.

• In the tropics the strongest atmospheric upwelling is over the warmest equatorial landmasses, causing three alternating highs and lows near the equator. This pattern can shift in the **El Niño** (or El Niño-Southern Oscillation (ENSO) phenomenon, where warm water moves from the vicinity of the great islands of southeastern Asia to the Central Pacific. Weather patterns may be drastically changed all over the globe during the ENSO.

• At higher latitudes the Coriolis effect inhibits direct transport of heat by the Hadley circulation; this is especially significant in winter when there is little direct solar input. Heat is instead transported by "eddy motions", involving the meandering of the **jet stream** and associated alternating cold outbreaks and winter storms. The jet stream is the result of very strong temperature contrasts between 30 and 50 N latitude.
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