

# 7: OCEAN-ATMOSPHERE INTERACTIONS



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### Critical Concepts

- **CC1 Density and Layering in Fluids:** Fluids, including the oceans and atmosphere, are arranged in layers sorted by their density. Air density can be reduced by increasing its temperature, and by increasing its concentration of water vapor causing the air to rise. This is the principal source of energy for Earth's weather.
- **CC3 Convection and Convection Cells:** Evaporation or warming at the sea surface decreases the density of the surface air mass and causes it to rise. The rising air mass cools by adiabatic expansion, and eventually loses water vapor by condensation, which increases temperature as latent heat is released. As air continues to rise, adiabatic expansion and radiative heat loss then cool the air and increase its density so that it sinks. These processes form the atmospheric convection cells that control Earth's climate.
- **CC4 Particle Size, Sinking, Deposition, and Resuspension:** Suspended particles (either in ocean water or in the atmosphere) sink at rates primarily determined by particle size: large particles sink faster than small particles. This applies to water droplets in the atmosphere. The smallest droplets sink very slowly and form the clouds, while larger droplets fall as rain.
- **CC5 Transfer and Storage of Heat by Water:** Water's high heat capacity allows large amounts of heat to be stored in the oceans and released to the atmosphere without much change of ocean water temperature. Water's high latent heat of vaporization allows large amounts of heat to be transferred to the atmosphere in water vapor and then transported elsewhere. Water's high latent heat of fusion allows ice to act as a heat buffer reducing climate extremes in high latitude regions.
- **CC8 Residence Time:** The residence time is the average length of time that molecules of contaminants, such as chlorofluorocarbons, spend in the atmosphere before being decomposed or removed in precipitation or dust. Long residence time allows such contaminants to diffuse upwards into the ozone layer in the upper atmosphere.
- **CC9 The Global Greenhouse Effect:** The oceans and atmosphere are both important in studies of the greenhouse effect, as heat, carbon dioxide and other greenhouse gases are exchanged between atmosphere and oceans at the sea surface. The oceans store large amounts of heat, and larger quantities of carbon dioxide both in solution and as carbonates.
- **CC10 Modeling:** The complex interactions between the oceans and atmosphere that control Earth's climate and affect the fate of greenhouse gases can best be studied by using mathematical models, many of which are extremely complex and require massive computing resources.
- **CC11 Chaos:** The nonlinear nature of ocean-atmosphere interactions makes at least part of this system behave in sometimes unpredictable ways and makes it possible for climate and ecological changes to occur in rapid, unpredictable jumps from one set of conditions to a completely different set of conditions.
- **CC12 The Coriolis Effect:** Water and air masses move freely over the Earth and ocean surface while objects on the Earth's surface, including the solid Earth itself, are constrained to move with the Earth in its rotation. This causes moving water or air masses to appear to follow curving paths across the Earth's surface. The apparent deflection, called the Coriolis Effect, is to the right in the Northern Hemisphere and to the left in the Southern Hemisphere. The effect is at a maximum at the poles, and is reduced at lower latitudes, becoming zero at the equator.
- **CC13 Geostrophic Flow:** Air and water masses flowing on horizontal pressure gradients are deflected by the Coriolis Effect until they flow across the gradient such that the pressure gradient force and Coriolis Effect are balanced, a condition called geostrophic flow. This causes ocean currents and winds to flow around high and low pressure regions in near circular paths. The familiar rotating cloud formations of weather systems seen on satellite images are formed by air masses flowing geostrophically.

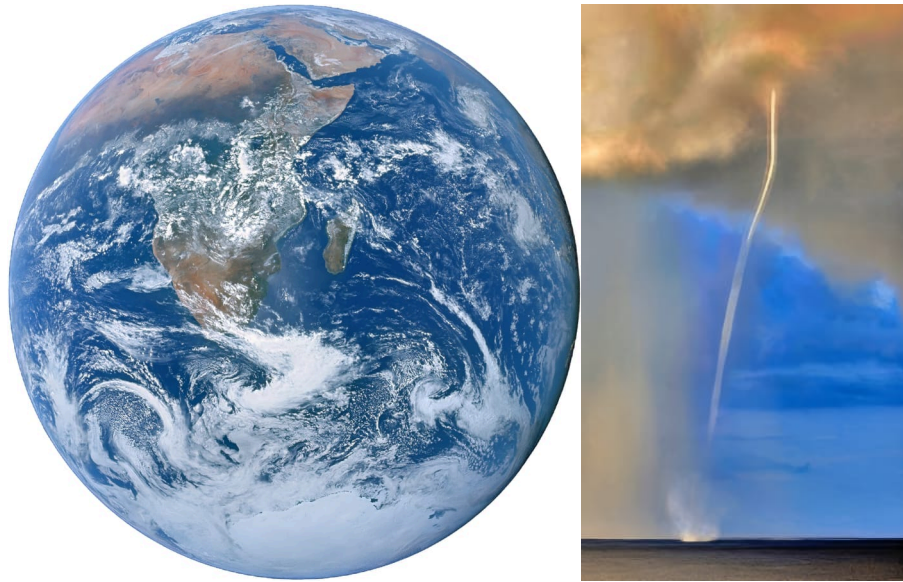


**Figure 7-1.** Atmospheric phenomena, including thunderstorms, waterspouts, hurricanes, and global winds and cloud patterns, are all driven in-part by heat energy released by ocean surface waters. (a) Apollo 17 took the first picture of Earth from space in 1972. The resulting picture, called the “blue marble” shows the swirling clouds and patterns of the land and oceans. (b) A waterspout, or tornado, is shown as it touches down in a strong thunderstorm near Grand Cayman Island in the Caribbean Sea.

The Earth’s **climate** (temperature, rainfall, etc., averaged over several decades of prior data) and **weather** (temperature, rainfall, etc., occurring at a specific time) are determined by the distribution of heat and water vapor in the atmosphere. The oceans play an important role in controlling this distribution.

Ocean-atmosphere interactions are also important because ocean **currents** are generated by winds as well as **density** differences between **water masses**. Seawater density is determined primarily by changes in temperature and **salinity** that occur at the ocean surface (**Chap. 5**). Surface water temperature is controlled by solar heating and radiative cooling, and salinity is altered by evaporation and precipitation.

The ocean contains many times more heat energy than the atmosphere does. This is because water is denser than air, and the total mass of ocean water is more than 200 times the total mass of air in the atmosphere. In addition, water has a much higher **heat capacity** per unit mass than air (or the rocks and soil of the land). Just the top few meters of ocean water have the same heat capacity as the entire atmosphere.

The oceans contain more than 97% of the world’s water. The ocean surface covers more than 70% of the Earth’s surface. The transfers of heat and water vapor between the ocean and the atmosphere that occur at the ocean surface are the main driving forces that determine the world’s climate.

Studies of the complex interactions between the oceans and atmosphere have been important since oceanographic science began. However, such studies have received more intensive attention because of potential climate changes as a result of the enhanced greenhouse effect (**CC9**). We cannot understand such climate changes without understanding ocean-atmosphere interactions.

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### 7.2: Water Vapor in the Atmosphere

### 7.3: Water and Heat Budgets

### 7.4: Climatic Winds

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## 7.1: The Atmosphere

Gases are highly compressible. Even small changes in pressure produce measurable changes in volume. As pressure increases, the gas molecules are forced closer together and the volume is reduced. In the atmosphere, pressure decreases rapidly with altitude (**Fig. 7-2**). Consequently, the density of the atmosphere decreases progressively with increasing height above the Earth's surface. As a result, the atmosphere is **stratified** throughout its more than 100-km height (**CC1**).

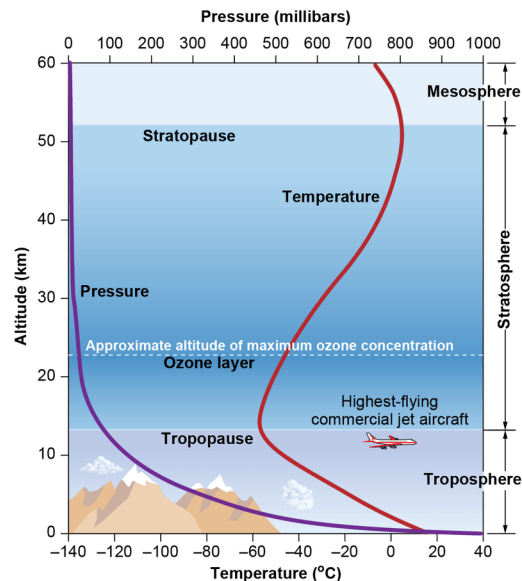


**Figure 7-2.** Structure of the atmosphere. The atmospheric circulation that determines the Earth's climate and weather takes place within the troposphere. Movements of gases between the troposphere and the stratosphere are limited and slow. Ozone is present throughout the stratosphere, but its concentration peaks at an altitude of about 22 km.

Temperature and vapor pressure (water vapor concentration) also affect the density of air, but much less than the changes in pressure with altitude. The atmosphere is steeply stratified near the Earth's surface because of the rapid pressure change with altitude. As a result, density driven vertical motions of air masses in the **troposphere** are limited to only a few kilometers above the Earth's surface.

### Atmospheric Structure

The atmosphere is separated vertically into three distinct zones: troposphere, stratosphere, and mesosphere. Beyond the mesosphere is a zone called the thermosphere which is a transition zone between the atmosphere and outer space. The troposphere lies between the Earth's surface and an altitude of about 16 to 18 km near the equator and less than 10 km over the poles (jet airplanes fly at an altitude of about 10 to 12 km). Vertical movements of air masses are caused by density changes and occur mainly in the troposphere. Although temperature generally decreases with altitude within the troposphere, density also decreases because the reduction in atmospheric pressure with altitude more than offsets the effect of lowered temperature (**Fig. 7-3**). This chapter pertains primarily to the troposphere because the atmospheric motions created by heat transfer from oceans to atmosphere are restricted to this zone.

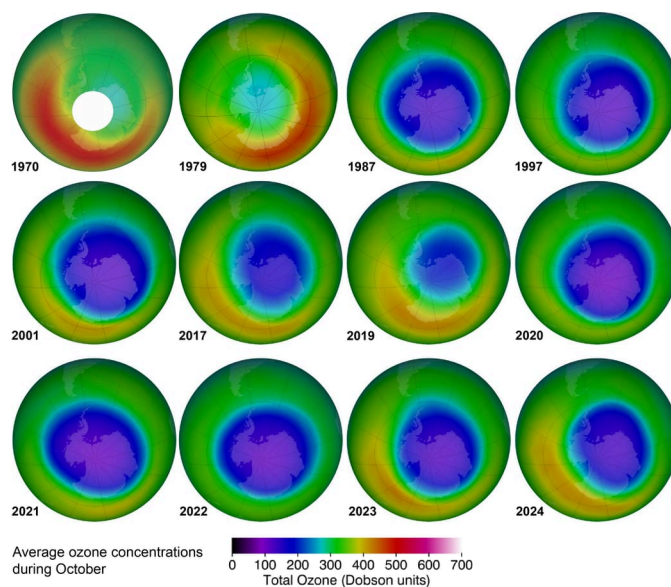


**Figure 7-3.** Average total amount of ozone above the Earth's surface over Antarctica during the month of October, as measured by a series of satellite instruments. The small circular area that has no color directly over the pole in the 1970 data is an area where no data were obtained. Nevertheless, it is clear that there is a broad region of very low ozone levels over Antarctica during the 1990s that was not present during the 1970s. This is the Antarctic ozone hole. The extent of the ozone hole has declined in recent years due to a global ban on the use of CFCs but other factors are likely involved. For example, the 2022 Hunga Tonga volcanic eruption may have contributed to an increase in ozone depletion in that year.

Between the top of the troposphere and an altitude of approximately 50 km, is the stratosphere. The air in the stratosphere is “thin” (that is, molecules are much farther apart than they are at ground level, because pressure is low), vertical air movements are primarily very slow **diffusion**, and temperature increases with altitude (**Fig. 7-3**). Within the stratosphere is a region called the **ozone layer**, where oxygen gas ( $O_2$ ) is partially converted to ozone ( $O_3$ ) by reactions driven by the sun's radiated energy.

## Ozone Depletion

The concentration of ozone in the Earth's ozone layer decreased progressively during the 1980s and 1990s, particularly in the region around the South Pole, where the depletion was so great that it has become known as the “ozone hole” (**Fig. 7-3**). The decrease varies seasonally for both poles and is greatest during the spring, which is in October for the Southern Hemisphere.

The ozone layer absorbs much of the sun's ultraviolet light, so depletion of the ozone layer could have severe adverse consequences for people and the **environment**. Ultraviolet light causes sunburn, eye cataracts, and skin cancers in humans; inhibits the growth of **phytoplankton** and possibly land plants; and may have harmful effects on other **species**.

The effects of ozone depletion on the environment are largely unknown. Studies have shown that the growth rate of phytoplankton in the Southern Ocean around Antarctica is significantly reduced (by 6 to 10%) in areas under the ozone hole. All life in the Antarctic, including whales, penguins, and seals, is ultimately dependent on phytoplankton for food (**Chaps. 12**). Consequently, besides the direct effects of the increased ultraviolet radiation, ozone depletion may adversely affect species by reducing their food supply.

There is still some uncertainty about the causes of ozone depletion, but the accepted view is that the primary cause is gases called **chlorofluorocarbons (CFCs)** released to the atmosphere. CFCs are synthetic chemicals used in many industrial applications. Until close to the end of the last century, most spray cans used CFCs to produce the aerosol and almost all refrigerators and air conditioners used substantial quantities of a mixture of CFCs called “freon.”

CFCs are highly resistant to decomposition and may remain in the atmosphere for decades. The long **residence time (CC8)** allows the CFCs to diffuse slowly upward until they reach the ozone layer. There they are decomposed by ultraviolet light, and their chlorine is released. The chlorine atoms take part in a complicated series of chemical reactions with ozone molecules whereby the ozone molecule is destroyed but the chlorine atom is not. Because it is not destroyed, the chlorine atom can destroy another ozone molecule and then another. In fact, one chlorine atom, on average, can destroy thousands of ozone molecules.

Ozone depletion in polar regions is greatest in spring because stratospheric clouds of supercold ice crystals are formed at that time. These clouds remove nitrogen compounds from the ozone layer. Thus, in spring, chlorine that would normally react with the nitrogen compounds reacts with the ozone instead.

CFC use as an aerosol gas in spray cans was banned and eliminated virtually worldwide in the early 1980s. Later, a global international treaty, the Montreal Protocol of 1987, was established that called for total elimination of CFC manufacture and replacement of these compounds by other chemicals by the year 2000. Although these measures will probably eliminate ozone depletion eventually, most scientists believe that the depletion will continue until at least the middle of this century, if not longer, until the CFCs already released have diffused slowly upward through the atmosphere and been decomposed. Also, the compounds used to replace CFCs destroy ozone in much the same manner as CFCs but far less effectively. The number of refrigeration units has rapidly increased and continues to increase as the standard of living rises globally. This has substantially delayed recovery of the ozone layer. Indeed, ozone depletion continues to be observed in both Northern and Southern Hemispheres although it appears to be decreasing slowly. However, other factors may interfere. For example, the 2022 Antarctic ozone hole was one of the largest on record, possibly partially due to the effects of the Hunga-Tonga volcanic eruption in January of that year. However, the year to year variability in the magnitude of the ozone hole is large so Hunga Tonga contribution is uncertain.

Depletion of ozone in the ozone layer in the stratosphere should not be confused with the problem of elevated ozone concentrations in smog in the troposphere (lower atmosphere). Ozone is released into the troposphere by various human activities and is created by photochemical reactions of other gases in the troposphere. Ozone is one of the principal harmful components of smog. However, this ozone does not contribute to the ozone layer because it does not last long enough in the atmosphere to be transported up through the troposphere and stratosphere to the ozone layer.

## 7.2: Water Vapor in the Atmosphere

In the troposphere, air masses move continuously, both vertically and horizontally, and these movements control the Earth's weather and climate and the winds that create ocean waves and ocean surface currents. The movements of air masses are caused primarily by changes in air mass density that occur as water vapor is added or removed.

### Water Vapor and Air Density

Water vapor in the atmosphere behaves in much the same way as sugar in water. Sugar dissolves in water, but the maximum quantity of sugar that can be dissolved is limited. More sugar can be dissolved at higher temperatures. Similarly, water vapor can be “dissolved” in air in limited quantities, and the amount of water vapor that can be present in air at saturation is greater at higher temperatures. The maximum amount of water that can remain in the vapor phase in air is expressed by the water vapor **saturation pressure**. The variation of water vapor saturation pressure with temperature is shown in [Figure 7-4](#). The saturation pressure of water vapor at average ocean surface temperatures (approximately 20°C) is many times greater than the saturation pressure of water vapor in air near the top of the troposphere (approximately -40°C to -60°C).

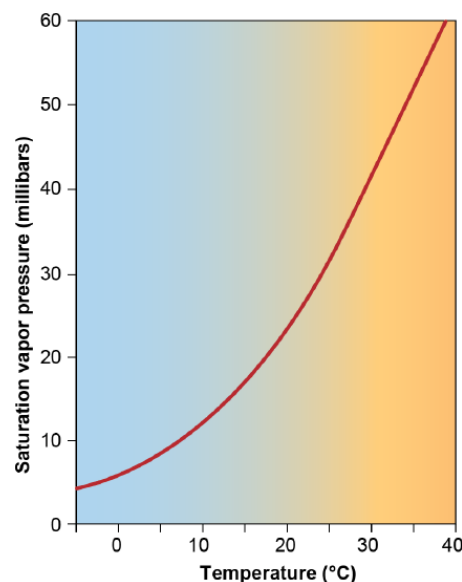


**Figure 7-4.** Variation of the water vapor saturation pressure with temperature in air. The water vapor saturation pressure is lower at lower temperatures. This is the reason that water condenses into rain as warm, moist air rises and cools.

When water evaporates from the sea surface, water molecules displace (move aside) oxygen and nitrogen molecules in the atmosphere. The displaced molecules are mixed and distributed in the surrounding air mass by random motions of the gas molecules. This is important because the atmospheric pressure and, therefore, density of the air mass would be increased if this were not the case. In contrast, water molecules ( $\text{H}_2\text{O}$ , molecular mass 18) are lighter than either the nitrogen ( $\text{N}_2$ , molecular mass 28) or the oxygen ( $\text{O}_2$ , molecular mass 32) molecules that they displace. Hence, light molecules displace heavier molecules, and the density of air is, therefore, reduced when water vapor is added (provided there is no change in pressure). As a result, moist air is less dense than dry air at the same temperature and pressure. Therefore, an air mass to which water vapor has been added tends to rise until it reaches its equilibrium density level in the density-stratified atmosphere (**CC1**). Conversely, if water vapor is removed, air increases in density and tends to sink.

Water vapor is continuously added to the atmosphere by evaporation of water from the ocean surface and removed by condensation and precipitation. The rate of evaporation varies with location, as explained later in this chapter. Where evaporation of ocean surface water is particularly rapid, the density of the air mass at the ocean surface is reduced, and the air tends to rise. In many parts of the oceans where the surface water is warmer than the overlying air, **conduction** and radiation of heat from the ocean surface also reduce the density of the air immediately above it.

## Water Vapor, Convection, and Condensation

If a gas expands without any external source of heat, its temperature decreases in a process called **adiabatic expansion**. As a result, when air rises in the atmosphere, the pressure decreases, and the air is cooled by adiabatic expansion. Because water vapor saturation pressure is highly temperature-dependent, cooling the air also results in a decrease in the water vapor saturation pressure. When the air has risen sufficiently that the actual vapor pressure becomes greater than the saturation vapor pressure, the air is supersaturated. In this condition, excess water vapor condenses into liquid water droplets (if above freezing). Because water molecules are typically too far apart to condense spontaneously, condensation in the atmosphere almost always occurs on cloud condensation nuclei (CCN).

In many cases, the water will condense to form large numbers of extremely small water droplets—droplets that are too small to fall out of the sky (**CC4**) or to combine with each other to form larger droplets. These water droplets form the clouds of the atmosphere. When cloud droplets are in the right environment for significant growth, precipitation can eventually form.

Water vapor contributes to vertical movements of and within air parcels, not only because of its low molecular weight, but also because of water's **latent heat of vaporization** (**Chap. 5**). When warm, moist air rises through the atmosphere and cools, and water vapor condenses, the water's latent heat of vaporization is released. The heat warms the air mass and thus reduces its density, causing the air parcels to rise. Release of latent heat is the major driving force for the spectacular rising plumes of moist air that form cumulonimbus (thunderstorm) clouds, which can reach the upper parts of the troposphere. Latent heat released to the atmosphere is also the main energy source for **hurricanes** and other storms.

Atmospheric **convection cells** are created by evaporation or warming at the sea surface that decreases the density of the air mass and causes it to rise. The rising air mass cools by adiabatic expansion, and eventually water vapor is lost by condensation. As the water vapor condenses, temperature is increased by the release of latent heat and the air tends to rise farther. Adiabatic expansion and radiative heat loss then cool the air parcel and increase its density so that it eventually sinks (**CC3**). The location, size, and intensity of convection cells determine the location and intensity of rainfall and snowfall, as well as the distribution of atmospheric heat energy. Convection cells exist at a range of spatial scales. We will discuss how global-scale cells form and how they influence climate and weather at different latitudes in the next several sections of this chapter.

## 7.3: Water and Heat Budgets

Averaged over the entire planet and over time periods of months or years, the amounts of heat and water that enter the atmosphere from all sources must equal the amounts removed from the atmosphere by all routes. If total inputs of heat or water did not equal total outputs, the average atmospheric temperature and/or the amount of water vapor in the atmosphere would progressively increase or decrease from year to year. In fact, the Earth's climate normally remains essentially unchanged for centuries.

In the Earth's past, very small imbalances between heat gained and heat lost have slightly changed the total amount of heat in the atmosphere and, thus, the Earth's climate. Today, **anthropogenic** releases of greenhouse gases, such as carbon dioxide, may disrupt, or already have disrupted, the balance between heat gained and heat lost that has remained stable over the past several centuries (CC9).

Because the inputs and outputs of heat and water to and from the atmosphere must balance, we can construct global heat and water budgets.

### Water Budget

The global water budget is relatively simple (Fig. 7-5). Water evaporates from the oceans, lakes, rivers, soils, and vegetation on land. The amount of water evaporated from land and ocean annually is enough to cover the entire globe with water almost 1 m deep. The evaporated water eventually condenses and falls as rain or snow over both land and ocean. Approximately 93% of the water that evaporates into the atmosphere comes from the oceans. However, only about 71% of the total global precipitation (rain and snow) falls on the oceans. Thus, more water evaporates from the ocean than reenters it as rain, and more rain falls on the land than is evaporated from it. This pattern is fortunate because the excess precipitation on land is the source of the Earth's freshwater supply. Excess water flows back into the oceans through rivers, streams, and groundwater.

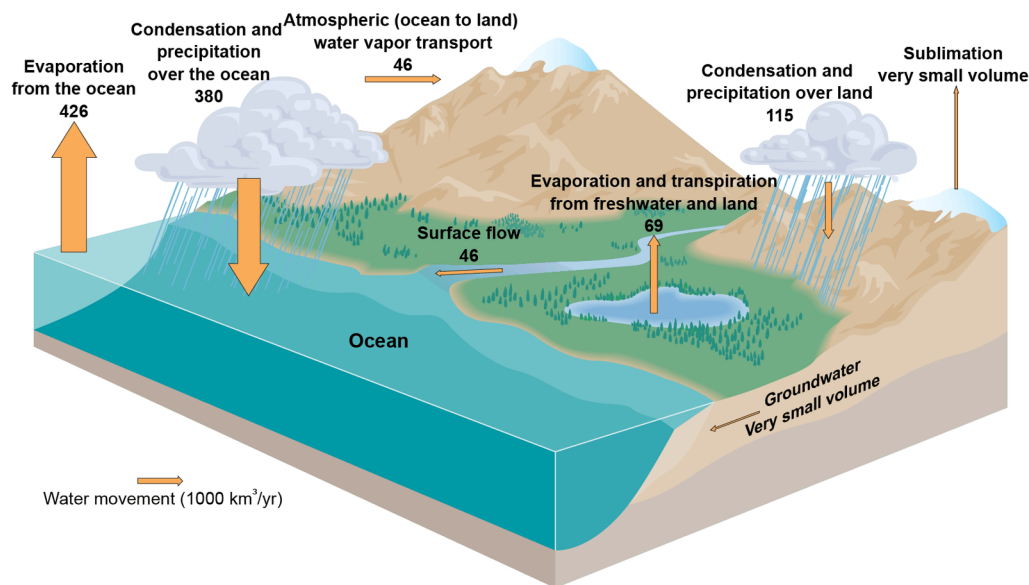


**Figure 7-5.** The Earth's water budget. The values represent the estimated volumes of water transferred annually between the various reservoirs by the indicated routes.

Neither the rate of water evaporation nor the amount of rainfall is the same everywhere on the oceans or on land. Clearly, deserts and tropical rainforests receive different amounts of rainfall and sustain different amounts of evaporation. The distribution of evaporation and rainfall is discussed later in this chapter.

## Heat Budget

The Earth's heat budget (Fig. 7-6) is more complicated than its water budget. The source of heat energy for the atmosphere is the sun. Solar radiation energy is partially absorbed by water vapor, dust, and clouds in the atmosphere and converted to heat energy. Part of the sun's energy is also reflected back, or backscattered, to space by clouds. Solar energy that reaches the Earth's surface is partially reflected and partially absorbed and converted to heat energy by the land and ocean waters. The Earth, ocean, atmosphere, and clouds also radiate heat energy. In fact, all bodies radiate heat. The peak wavelength radiated increases as the temperature of the body decreases. Because the Earth is much cooler than the sun, the wavelengths radiated by the Earth, ocean, and clouds (infrared) are much longer than the wavelengths emitted by the sun (visible and ultraviolet). The difference in wavelength and the differential absorption by atmospheric gases at different wavelengths are the basis for the greenhouse effect (CC9).

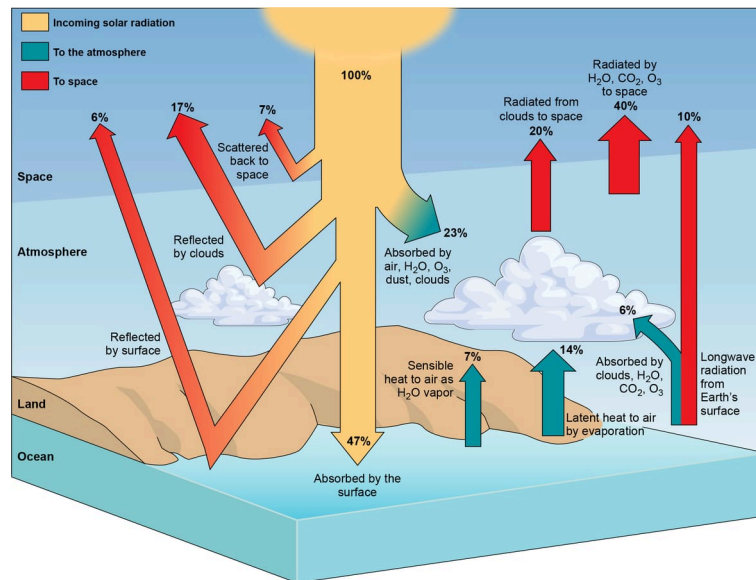


**Figure 7-6.** The Earth's heat budget. Solar radiation, which is concentrated in the visible light wavelengths, is reflected, scattered, and absorbed by the gases and clouds of the atmosphere and by the ocean and land surface. All of the absorbed energy is radiated back to space, primarily as infrared radiation. The diagram shows the estimated percentages of the incoming solar radiation that are absorbed, reflected, and reradiated to space by the various components of the Earth-ocean-atmosphere system. Each of these may change in an unpredictable way as the greenhouse effect is enhanced by anthropogenic inputs to the atmosphere (CC9, CC10, CC11).

About 20 to 25% of the solar radiation reaching the Earth is absorbed by atmospheric gases and clouds and converted to heat energy. Almost 50% of the solar radiation is absorbed by the ocean water and the land, then transferred to the atmosphere by the conduction of **sensible heat**, as latent heat of vaporization, and by radiation. The net transfer averaged over the global oceans is from ocean to atmosphere. Heat transfer by conduction is relatively small, but the transfer of latent heat from ocean to atmosphere equals about one-quarter of the total solar energy that reaches the Earth. Thus, the oceans effectively capture a major portion of the sun's radiated energy and transfer much of it to the atmosphere as latent heat of vaporization. This mechanism moderates climate differences between Earth's tropical and polar regions.

## Latitudinal Imbalance in the Earth's Radiation

At the equator, the sun passes directly overhead at noon on the spring and autumnal **equinoxes** (Fig. 7-7a). The sun's radiated energy therefore passes vertically through the atmosphere, and the **angle of incidence** at the Earth's surface is 90°. At the poles on the equinoxes, the sun remains on the horizon all day, and its radiation must pass through the atmosphere parallel to the Earth's surface.

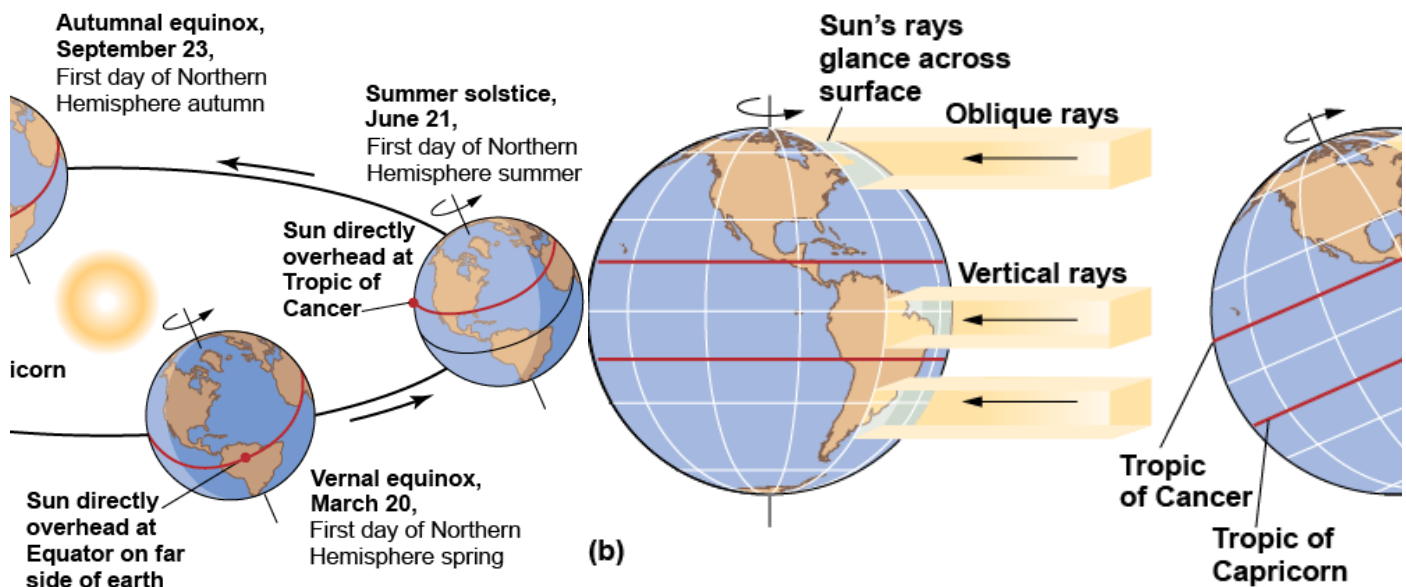


**Figure 7-7.** (a) The Earth, spinning on its tilted axis, moves around the sun once a year. In northern summer (on the right), the Earth is tilted such that the sun is directly overhead at a latitude north of the equator at a point that moves daily around the Earth. Similarly, in northern winter, this point is south of the equator. At the spring and autumnal equinoxes, the sun is overhead at the equator throughout the day as the Earth spins. (b) The angle of incidence of solar radiation received at the Earth's surface also changes with the seasons, but generally, the area over which a given amount of the sun's energy is spread is less near the equator and increases toward the poles. The globe on the left shows the Earth at an equinox. The globe on the right shows the Earth at the winter (northern) solstice.

Because the poles are only about 6000 km farther from the sun than the equator is, and because the mean distance from the sun to the Earth is 148,000,000 km, the intensity of the sun's radiated energy is only about 0.01% lower at the poles than at the equator. However, solar energy is spread over a progressively larger area with increasing **latitude** (Fig. 7-7b). Consequently, on the equinoxes, the solar radiation received by an area of the Earth's surface at 30°N or 30°S is only 86% of that received by an equal area at the equator. The proportion at 60°N or 60°S is only 50%, and at the poles it is zero.

Nearer the poles, solar radiation must also travel a longer path through the atmosphere because of the lower angle to the Earth's surface. More of the solar radiation is absorbed as it passes through the atmosphere because of the increased path length, but this effect causes a relatively small difference in solar intensity at the Earth's surface.

The intensity of the Earth's long-wavelength radiation increases with temperature. However, the temperature difference between equatorial and polar regions produces a latitudinal variation in the Earth's longwave radiation intensity that is not as great as the latitudinal variation of solar insolation of Earth's surface (Fig. 7-8).

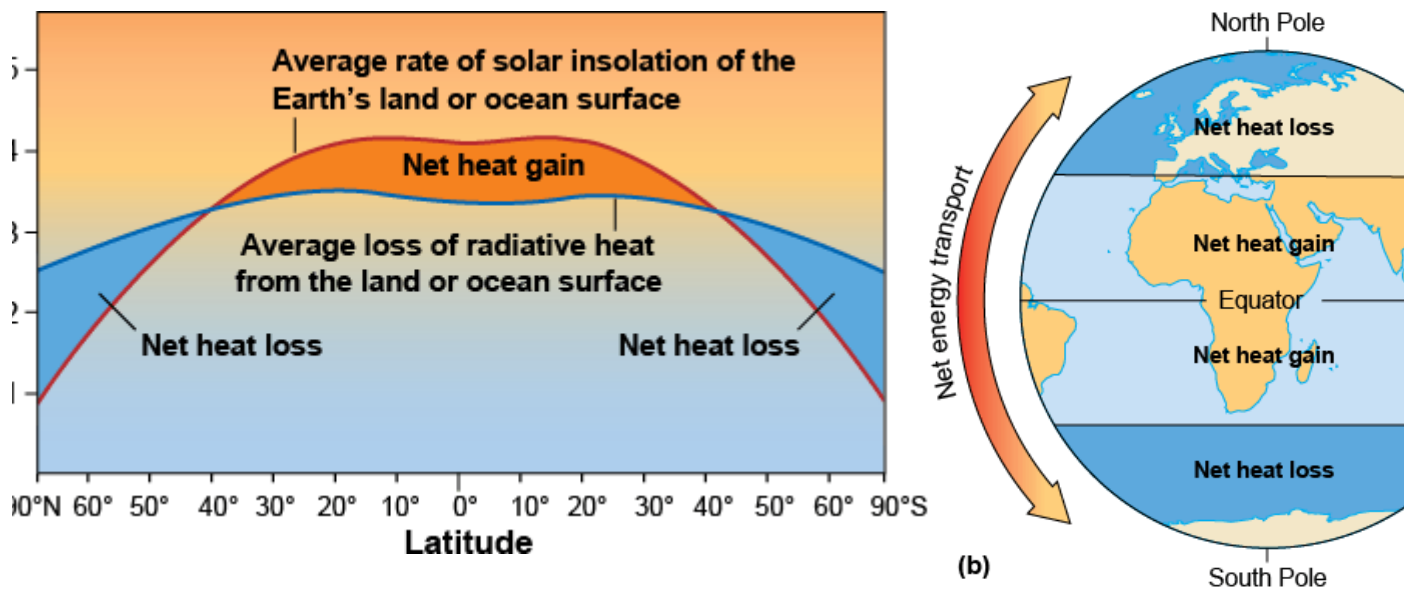


**Figure 7-8.** Variation in radiative heat loss and gain at the Earth's surface with latitude. There is a net gain between the equator and about 35° to 40°, and a net loss at higher latitudes.

The total solar radiation absorbed by the Earth must equal the total long-wavelength radiation lost by the Earth when averaged over the whole Earth, but this balance does not occur at all latitudes. There is a large difference between the rates at which solar heat is absorbed at the poles and at the equator. There is a much smaller difference between the rates at which long-wavelength radiation is lost at the poles and at the equator ([Fig. 7-8](#)). Consequently, there must be mechanisms for transferring heat energy from low latitudes to high latitudes. If heat were not transferred, the polar regions would be much colder and the equatorial regions much warmer. Planets and moons that have no (or a limited) mechanism of heat transfer between latitudes have dramatically greater differences in surface temperature between their polar and equatorial regions than does the Earth. Hence, latitudinal heat transfer is one of the critical mechanisms that maintains the habitability of the Earth's entire surface.

Heat is transported latitudinally in two ways. First, latent heat of vaporization enters the atmosphere as water is evaporated in warm tropical regions and some of the heat is transported to higher latitudes by atmospheric convection cells. At higher latitudes, cooling causes water vapor to condense in the form of rain or snow, thus releasing latent heat to the atmosphere. Second, ocean water warmed by the sun in low latitudes is transported by ocean currents to higher latitudes, where it transfers heat to the atmosphere by conduction and evaporation. Near the equator, ocean currents transport more heat to higher latitudes than does atmospheric transport, whereas near the poles atmospheric transport of heat to higher latitudes is greater than ocean current transport.

## 7.4: Climatic Winds

Winds are extremely variable from one day to the next at any location. The day-to-day variations are part of what we call “weather.” Weather events may include calm, thunderstorms, hurricanes, fog, and the other phenomena that a weather forecaster tries to predict hours or days before they happen. If we average the day-to-day variation over relatively large areas and over periods of several days or weeks, we find patterns of average winds, temperatures, rainfall, and so on that are relatively repeatable from one year to the next in each area. These long-term average patterns are called “climate.” This chapter is concerned primarily with climatic winds, temperatures, and rainfall, although some aspects of various weather events are also examined.

### Climatic Winds on a Nonrotating Earth

On a nonrotating Earth with no continents or **Coriolis effect (CC12)**, wind patterns would be simple (**Fig. 7-9**). More heat is transferred to the atmosphere from the oceans than is absorbed by the atmosphere directly from the sun (**Fig. 7-6**). Solar heating of the oceans is greatest at the equator. Thus, ocean water is warmer and the rate of heat transfer by evaporation of water from ocean to atmosphere is greater than at higher latitudes. The equatorial air mass would rise through the troposphere, creating a low-pressure zone and horizontal **pressure gradient** in the atmosphere at sea level. The rising air would be replaced by air flowing into the region from higher latitudes. Once it reached its equilibrium density level (**CC1**), the rising air would spread toward the poles. As the air mass moved north or south in the upper troposphere, it would progressively cool and release its water vapor as rain or snow.

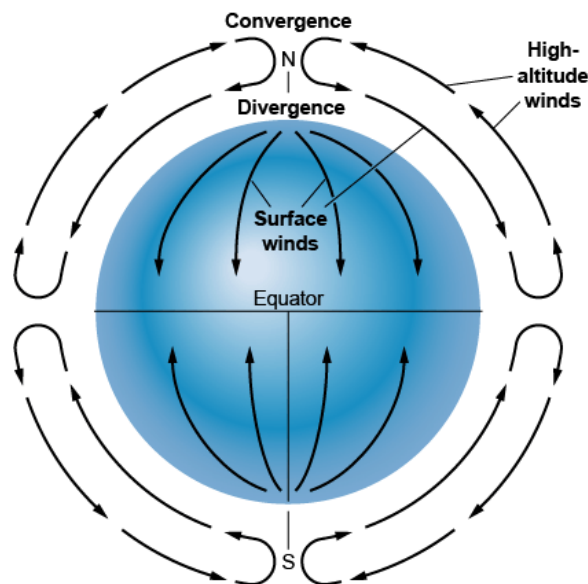


**Figure 7-9.** Atmospheric convection cells and winds on a hypothetical water-covered, nonrotating Earth.

Once the air mass reached the pole, it would have sufficiently cooled and lost water vapor by condensation or deposition to become dense enough to sink and create a high-pressure zone in the atmosphere at sea level. The air would then move toward the equator along the horizontal atmospheric pressure gradient between the polar high-pressure zone and the equatorial low-pressure zone. As the air moved back toward the equator, it would be warmed and would gain water vapor from evaporation of ocean surface water. This is a simple convection cell motion (**CC3**). There would be a **convergence** at sea level at the equator, and **divergence** at sea level at the poles. Surface winds would blow from the poles toward the equator in each hemisphere. This simple two-convection-cell pattern (**Fig. 7-8**) illustrates the basic principles of atmospheric circulation, but it does not exist on the Earth, because we have so far not accounted for the fact that the planet rotates and moving air is deflected by the Coriolis Effect. The presence of land is also not yet accounted for, but we can best understand the complications that land introduces if we first look at a hypothetical water-covered Earth that is rotating.

## Climatic Winds and the Coriolis Effect

The deflection of air movements by the Coriolis Effect, caused by Earth's rotation, results in an atmospheric circulation system that consists of three **convection** cells arranged latitudinally in each hemisphere (**Fig. 7-10**). To some extent, this real-world system has the same net result as the simple two-convection cell system: air rises at the equator and sinks at the poles, and heat is transported from the equator to the poles.

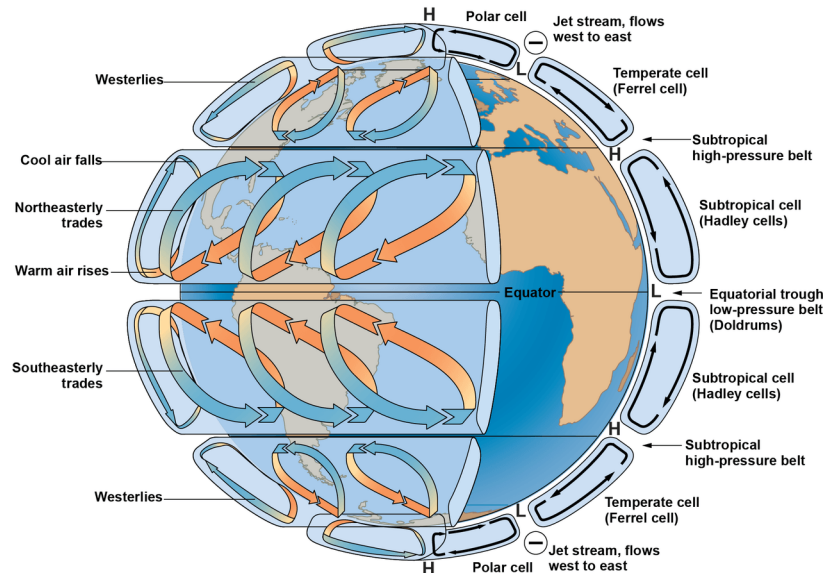


**Figure 7-10.** Atmospheric circulation cells and winds on a hypothetical water-covered, rotating Earth are arranged in latitudinal bands. There are three cells in each hemisphere. Upwelling occurs at the equator and the polar fronts, and downwelling at the poles and in mid latitudes. Note how the Coriolis effect deflects moving air masses: air masses moving away from the equator are deflected to the east, and air masses moving toward the equator are deflected to the west. Atmospheric convection cells are arranged in this general pattern on the Earth, but they are modified substantially by the influence of the land masses.

To understand how a three-cell per hemisphere circulation pattern develops, consider the movements of an air mass initially located at sea level on or near the equator. The air mass is heated by solar radiation and has a high water vapor content because of the relatively high evaporation rate from the ocean's surface. It rises through the troposphere to its density equilibrium level and spreads out toward the poles, just as it would on a nonrotating Earth. As the air mass moves away from the equator, where there is no deflection by the Coriolis Effect, it is not deflected significantly until it reaches  $5^{\circ}$  to  $10^{\circ}$ N or S (**Fig. 7-10**). As it continues to move toward the poles, the air mass is deflected increasingly toward the east because the Coriolis Effect increases with latitude.

At about  $30^{\circ}$ N or  $30^{\circ}$ S, the now eastward-moving air has risen, expanded, cooled, and lost most of its water vapor. The cooler, dry air sinks to the surface, where some air moves toward the pole and some moves away from it, forming a high-pressure zone and atmospheric divergence at ground level. Air that moves toward the pole in this region enters the "Ferrel cell" circulation (**Fig. 7-10**). Air that moves back toward the equator is deflected toward the west. The deflection decreases as the air mass moves nearer the equator. Consequently, surface winds in the subtropical regions blow predominantly from the northeast in the Northern Hemisphere and from the southeast in the Southern Hemisphere (**Fig. 7-10**). These are the **trade winds**, and the convection cell air movements that occur between the equator and about  $30^{\circ}$ N and  $30^{\circ}$ S are called "Hadley cells."

Surface winds in the Ferrel cells (**Fig. 7-10**) are also deflected by the Coriolis effect, but to the east because they are moving away from the equator. Therefore, this is a region of persistent surface winds from the southwest in the Northern Hemisphere and from the northwest in the Southern Hemisphere, commonly called the **westerlies**. Winds are named for the direction from which they are blowing from.

Although the predominant winds are from the east in subpolar regions, air movements in the polar cells are complex because of the stronger Coriolis deflection and the presence of the **jet streams**. Jet streams are swiftly moving west-to-east air currents high in the troposphere. They are located over the boundaries between the Ferrel and polar cells, called the "polar fronts" (or "Antarctic front" in the Southern Hemisphere). The jet streams undergo complex **meanders** associated with movements of the polar fronts. Because storms often form along the jet stream, and because meanders in the jet stream can bring cold polar air to lower latitudes and warm air to higher latitudes, the orientation and trajectory of the jet stream are routinely reported in many

weather forecasts.

In any one region, the predominant wind direction and its persistence, the average extent of cloud cover, and the average rainfall are determined largely by the region's location with respect to the atmospheric convection cells (**Fig. 7-10**). In the equatorial region known as the Doldrums, where trade winds converge and atmospheric upwelling occurs, sea-level winds are light and variable. In addition, because the rising moist air is cooled, cloud cover is persistent and rainfall is high. This region is called the **intertropical convergence zone** because it is where two wind systems converge. Air rising from the sea surface through the troposphere causes low atmospheric pressure at sea level in the intertropical convergence zone. Air also rises through the troposphere at the polar front. However, unlike the intertropical convergence zone that migrates seasonally but otherwise remains generally invariable, the polar front oscillates in wavelike motions, called "Rossby waves" (**Chap. 9**), which have timescales of days or weeks. These waves can spawn massive storms called **extratropical cyclones**.

In the atmospheric **downwelling** (or subsidence) regions between the Hadley cells and Ferrel cells and at the poles, atmospheric pressure is high, winds are light, skies are usually cloudless, and rainfall is low. The **subtropical high**-pressure zones between the Hadley and Ferrel cells are called the "horse latitudes." Sailing ships were often becalmed for long periods in these regions, and the horses on board were killed to conserve water. Because of the high-pressure zone at the South Pole, much of Antarctica is a desert with very low annual precipitation. In fact, there are dry desert valleys in Antarctica with no snow cover, and the interior of Antarctica receives only about 50 mm of precipitation per year (desert climates are defined as having less than 254 mm of rain per year). The massive amounts of ice and snow in the thick ice sheet that covers other parts of Antarctica took millennia to accumulate, and the ice sheet exists only because Antarctic temperatures are too low for snow to melt during the summer.

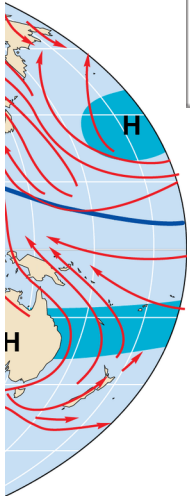
Under the centers of the atmospheric cells (between convergences and divergences), winds are generally persistent, particularly in the trade wind zones. Rainfall and cloud cover are variable, and the zones are affected by periodic strong storms, especially in the higher-latitude areas of the westerly wind zones, where storms form on the polar front. Westerly surface winds under the Ferrel cells can be particularly strong, especially in the Southern Hemisphere, where few continents disturb the circulation. Sailors call the area of strong westerly winds at about 40°S the "roaring forties."

The atmospheric convection cell system described here is a simplified view of the climatic wind patterns on the Earth's surface. Complications of this simplified pattern occur because the interactions between land and atmosphere differ from the interactions between ocean and atmosphere. The pattern of high- and low-pressure zones and winds described by the six-cell convection system model generally holds true over the oceans but is much modified over the large land masses of the continents. The strong Coriolis Effect and shallow troposphere of the polar cell regions favor horizontal transport of air masses by weather systems and poor development of vertical convection cell motion. Complications in the six-cell system also occur because of the seasonal movement of the Earth around the sun and certain longer-term oscillations of the ocean-atmosphere interaction, such as **El Niño**, which is described later in this chapter. Finally, local wind patterns are highly variable because weather events operate at much smaller geographic scales than do the global climatic wind patterns discussed in this section.

## Seasonal Variations

Because the Earth's axis is tilted in relation to the plane of the Earth's orbit around the sun, the latitude at which the sun is directly overhead at noon changes progressively during the year. At the Northern Hemisphere summer **solstice** (June 20 or 21), the sun is directly overhead at 23.5°N, the Tropic of Cancer. At the autumnal equinox (September 22 or 23) and the spring equinox (March 20 or 21), the sun is directly overhead at the equator. At the Northern Hemisphere winter solstice (December 21 or 22), the sun is directly overhead at 23.5°S, the Tropic of Capricorn (**Fig. 7-8a**).

As the Earth moves around the sun, the latitude of greatest solar intensity migrates north and south, resulting in a corresponding northward or southward displacement of the atmospheric convection cells. The displacement of atmospheric convection cells causes seasonal climate variations. The seasonal movement of convection cells is shown by a seasonal shift in wind bands (**Fig. 7-11**) and in surface-level atmospheric high- and low-pressure zones (**Fig. 7-12**).

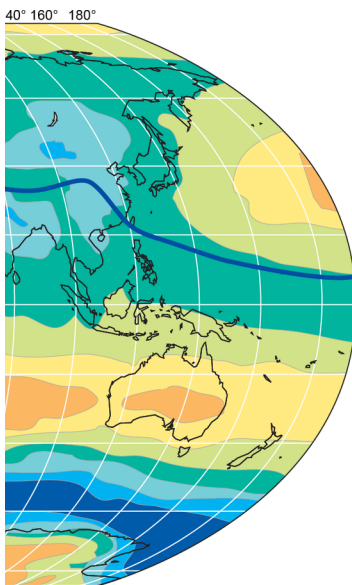


Subtropical high



Prev

**Figure 7-11.** Climatic winds over the ocean surface in (a) January and (b) July. Winds generally blow from higher pressure toward lower pressure. Note the seasonal migration of the subtropical high-pressure zones, the intertropical convergence zone, and the associated trade winds. The migration is more pronounced over the major continental landmasses than over the oceans. Over the Indian Ocean, the seasonal migration is enhanced by the strong monsoonal interaction between Asia and the Indian Ocean.



Intertropical converge

**Figure 7-12.** Mean atmospheric pressures at sea level in (a) January and (b) July. The greater complexity of the geographical distribution and seasonal change of atmospheric pressure in the Northern Hemisphere is due to the concentration of the landmasses in this hemisphere. The ocean moderates regional coastal climates, leading to larger seasonal changes in pressure over large landmasses such as Asia.

Because the latitude at which tropospheric air temperature is highest changes seasonally, the convection cells also migrate seasonally. Tropical tropospheric air temperature is controlled primarily by radiation and evaporative heat inputs from the oceans and land. Because water has a high heat capacity, a great amount of heat energy is necessary to warm the ocean surface waters (CC5). Therefore, water temperature, evaporation rate, and air temperature do not increase in step with changes in solar intensity as the latitude of greatest solar intensity shifts north and south. Several weeks of increased (or decreased) solar heating must elapse before these parameters change enough to cause the atmospheric convection cells to move. Consequently, there is a time lag between the sun's seasonal movement and the movement of the atmospheric convection cells. The time lag is why many northern countries have their warmest weather in August and their coldest weather in February, approximately 2 months after the summer and winter solstices occur in June and December, respectively.

Heat captured from solar radiation is transferred to the atmosphere more rapidly by land than by oceans, primarily because of the high heat capacity of water. As a result, the time lag between seasonal changes in solar radiation and air temperature is

shorter over the continents than over the oceans. Because the continents are concentrated in the Northern Hemisphere, seasonal migration of convection cells has a shorter time lag north of the equator than south of the equator. Furthermore, the center of the latitudinal wind bands, the intertropical convergence zone, migrates farther north of the equator in the Northern Hemisphere summer (especially over the continents) than it does south of the equator during the Southern Hemisphere summer (**Fig. 7-12**).

Seasonal changes in climate are greater in areas that are under an atmospheric convergence or divergence zone in one season and under the middle of a convection cell in another. Seasonal climate changes are also greater in higher latitudes than near the equator. The reason is that the total amount of solar heat reaching the Earth's surface per day depends on both the sun's angle and the amount of time during which the sun is above the horizon on a given day. Both the sun's intensity and the day length vary more between summer and winter at high latitudes than at low latitudes.

## Monsoons

In the Northern Hemisphere summer, the intertropical convergence zone moves farther north over the landmasses of Asia and Africa than it does over the oceans. Warm, moist tropical air moves north onto Asia from the Indian and western Pacific Oceans in the trade wind zone under the southern Hadley cell (**Fig. 7-13a**). Thus, in summer the southern Hadley cell trade winds extend into the Northern Hemisphere, where the Coriolis deflection is to the right. These winds, called monsoon winds, are deflected by the Coriolis Effect in the Northern Hemisphere, so they blow from the southwest, whereas trade winds in the Southern Hemisphere blow from the southeast.

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### Intertropical convergence zone Prev

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**Figure 7-13.** Indian Ocean monsoon winds reverse seasonally. (a) In summer, an intense low-pressure zone forms over Asia as the land is heated by the sun and loses this heat to the atmosphere more quickly than does the ocean. The heated air rises to form the low-pressure zone at a latitude where, if the Earth were covered by ocean, the Hadley and Ferrel cells would intersect and atmospheric pressure would be high. The low-pressure zone causes the atmospheric convection cells to shift to the north and bring warm, wet, monsoon winds from the southwest to parts of Asia. A similar, but less pronounced effect occurs over North Africa. (b) In winter, the Asian landmass loses heat more quickly than does the ocean, and an intense high-pressure zone is formed by the sinking cold, dry air mass. As this air mass spreads to the south, it shifts the convection cells south, which creates the northeast monsoons and brings dry weather to much of Asia.

The warm, moist air carried by monsoon winds causes storms and torrential rains in India and Southeast Asia. This densely populated region depends on the rains to support agriculture. However, the area is plagued by alternating floods and droughts. When monsoons are strong, devastating floods occur. When monsoons are weak or absent for a year or two, drought and famine occur. The nature of the multiyear, or “interannual,” climate variations that cause prolonged famine or drought is just beginning to be understood. We discuss some of these variations later in this chapter.

In the northern winter, the intertropical convergence zone moves south over the Indian and Pacific Oceans and Africa. In addition, the divergence between the Hadley and Ferrel cells moves far to the south over Asia in winter for the same reason that the intertropical convergence zone moves far north in summer. Thus, in winter this downwelling zone lies partly south of the Himalayan Mountains. The cool, dry downwelling air flows south over India, producing dry winters (**Fig. 7-13b**). Similar, but weaker and more complicated, seasonal monsoon wind reversals occur throughout East Asia, tropical Africa, and northern Australia (**Fig. 7-13**).

Sailors have taken advantage of the monsoons for thousands of years. Coastal traders sail on winter trade winds from India and Southeast Asia to East Africa and Madagascar and returned on the summer monsoon winds.

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## 7.5: Climate and Ocean Surface Water Properties

Ocean surface water salinity and temperature are controlled by solar radiation, transfer of heat and water between atmosphere and oceans, ocean currents, vertical mixing, and locally, river runoff.

### Ocean Surface Temperatures

**Figure 7-8** shows the generalized latitudinal distribution of solar radiation that reaches the Earth's surface. Ocean surface water temperatures generally decrease from the equator toward the poles (**Figs. 7-14, 7-18**), and thus, they reflect the distribution of solar radiation.

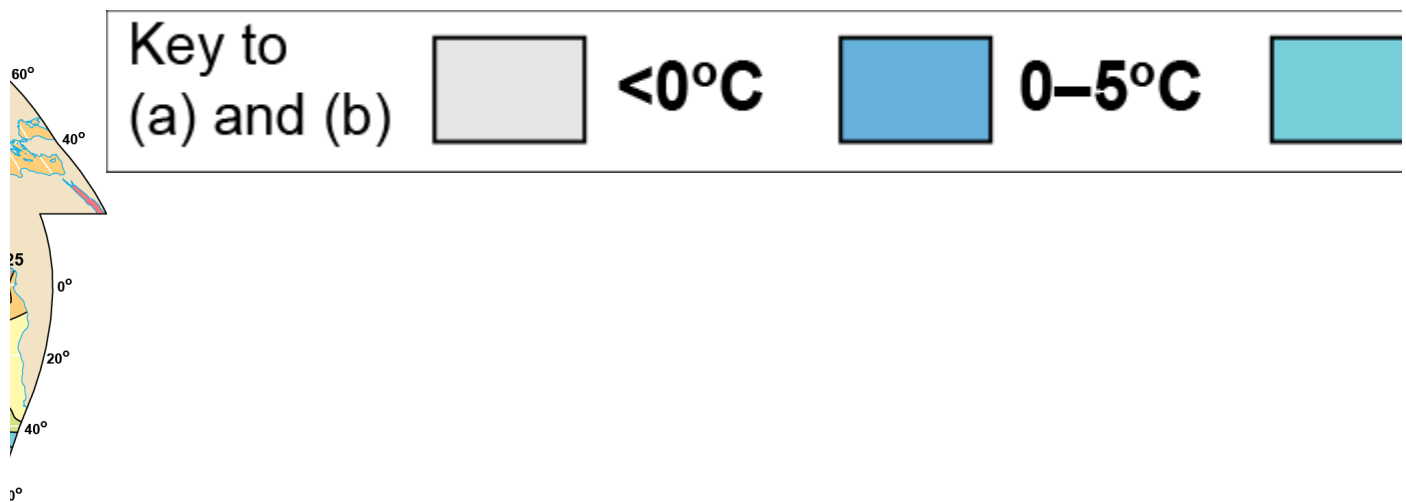


**Figure 7-14.** Average surface water temperatures in the world oceans in (a) February and (b) August.

On an Earth without ocean currents or winds, ocean surface water temperatures would decrease uniformly with latitude, and the **isotherms** in **Figure 7-14** would be horizontal lines parallel to lines of latitude. Although such a basic pattern is evident in **Figure 7-14**, complications are also readily apparent.

First, a band of low-temperature water lies along the equator on the eastern side of the Pacific Ocean. The band is a surface water divergence where cold, deep water is continuously upwelled (**Chap. 8**). The relatively low surface temperature at the equator in the Atlantic Ocean may be a result of similar, but less well-developed, upwelling. In the Indian Ocean, upwelling at the equator is inhibited by the monsoon winds.

The second complication evident in **Figure 7-14** is that isotherms in mid latitudes are inclined toward higher latitudes on the western sides of the oceans. Such a pattern is especially evident in the North Pacific and North Atlantic Oceans. It is a result of warm surface ocean currents that flow poleward from the subtropics on the western sides of the ocean basins and then across the ocean to the east. The currents are part of the **subtropical** gyres discussed in **Chapter 8**. In the North Atlantic Ocean the current is the Gulf Stream, and in the North Pacific Ocean it is the Kuroshio Current.

Ocean surface water temperature responds to seasonal changes in the sun's angle. High temperatures typical of tropical and subtropical latitudes extend to higher latitudes in summer (**Fig. 7-14b**). However, because of the heat-buffering action of ice (**Chap. 5**), surface water temperatures do not vary significantly with the seasons in the Arctic Ocean or the ocean surrounding Antarctica.

## Ocean Surface Salinity

Differences in the salinity of ocean surface waters are caused by variations in the rate of ocean surface water evaporation, by variations in the rate of freshwater input from rainfall and land runoff, and, in limited areas, by upwelling and downwelling. Because geographic variations in evaporation and precipitation rates are somewhat complex, the distribution of ocean surface water salinity (**Fig. 7-15**) is somewhat more complicated than that of surface temperatures. Salinity is generally highest in surface waters in subtropical regions that are remote from land. It is generally lower at the equator, in polar and subpolar regions, and near continents. Surface salinity is higher in the subtropical Atlantic Ocean than in the subtropical regions of the other oceans.

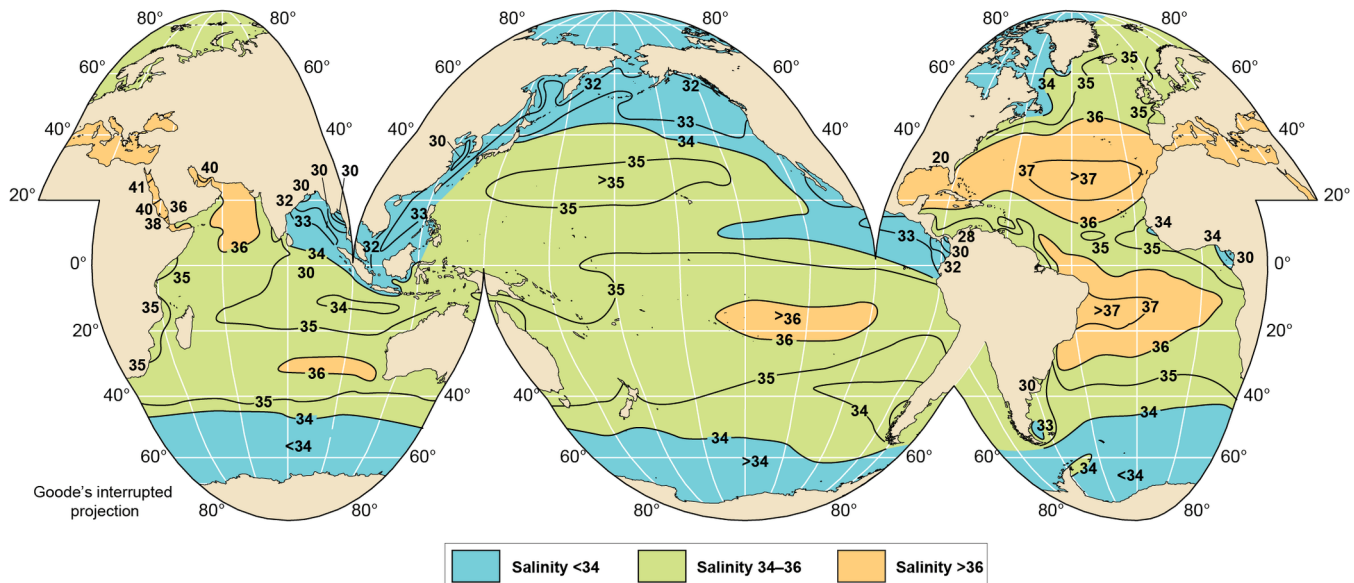


**Figure 7-15.** Surface water salinity in the world oceans.

The distribution of salinity is related to variations in evaporation and precipitation with latitude (**Fig. 7-16**). Precipitation is high near the equator and at mid latitudes around 40° to 60°N and S, which are the atmospheric convergence regions between the Northern and Southern Hemisphere Hadley cells and between the polar and Ferrel cells (**Figs. 7-10, 7-11**). In these regions, warm, moist air converges and rises through the atmosphere. As the air rises, its water condenses to form clouds and precipitation. Precipitation is lower in subtropical latitudes (the horse latitudes) and near the poles because these regions are under the centers of the atmospheric convection cells or at divergences.

Evaporation varies with latitude in a different way than precipitation does (**Fig. 7-16**). The evaporation rate is higher at warmer ocean surface water temperatures. Hence, evaporation generally decreases progressively from equator to pole. However, the evaporation rate is lower in equatorial latitudes than in subequatorial regions for two main reasons. First, persistent cloud cover reduces the solar intensity in the equatorial region. Second, the equatorial region is persistently calm, whereas the trade wind regions have higher winds and hence greater evaporation. The higher rate of evaporation is due to both increased airflow over the water and the increased surface area of the water caused by waves, particularly breaking waves, which create water droplets and bubbles with relatively large surface areas.

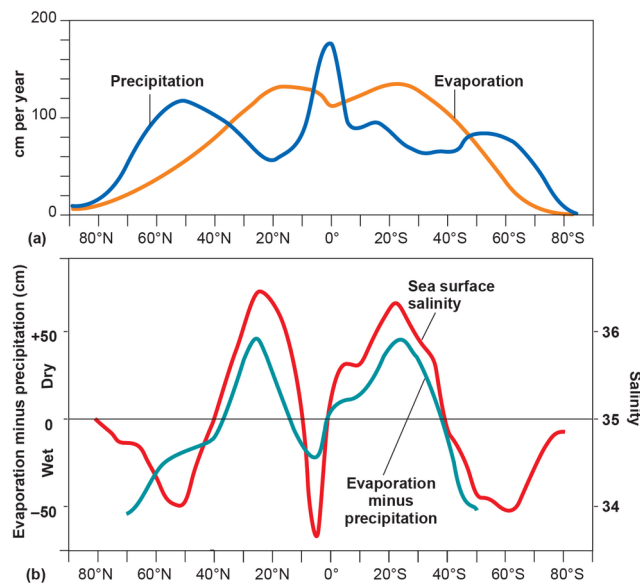


**Figure 7-16.** Precipitation and evaporation as a function of latitude. (a) Precipitation is high at the atmospheric upwelling zones at the equator and at 50° to 60°N and S, somewhat lower in mid latitudes, and very low at high latitudes. Evaporation generally increases from the poles toward the equator, but it is slightly lower at the equator because of the extensive cloud cover. Thus, there is a net excess of evaporation over precipitation in mid latitudes, whereas there is a net excess of precipitation within a few degrees north and south of the equator and at latitudes above approximately 40° in both hemispheres. (b) The salinity of surface ocean waters varies according to the difference between evaporation and precipitation. It is high where evaporation exceeds precipitation, and low where precipitation exceeds evaporation.

**Figure 7-16b** shows that precipitation exceeds evaporation across the equatorial oceans and at latitudes above about 40°N and 40°S. Therefore, surface salinity tends to be lower in these regions. In contrast, evaporation exceeds precipitation in subtropical regions. Consequently, surface salinity tends to be higher in these regions (**Fig. 7-16b**). The pattern of higher salinity in subtropical regions and lower salinity in equatorial regions and at high latitudes is observed in surface waters of the central parts of each ocean (**Fig. 7-15**). The distribution is more complex in some regions near continents because of large quantities of freshwater runoff.

The distribution of precipitation with latitude is substantially altered by the presence of landmasses (**Fig. 7-17**). The higher surface salinity in the central North and South Atlantic oceans (**Fig. 7-15**) is the result of interactions of the mountain chains of North and South America with the prevailing winds, and the outflow of high-salinity water from the Mediterranean (**Chap. 8**). In the westerly wind zones, moist air masses that move from the Pacific Ocean onto the American continents lose their moisture as rainfall on the west side of the Andes in South America and on the west side of the Rocky Mountains, the Sierra Nevada, and the coastal mountain ranges in North America. Consequently, the precipitation runs off into the Pacific Ocean. In contrast, the Northern Hemisphere trade winds in the Atlantic Ocean blow across the relatively narrow, low-altitude neck of Central America and therefore carry their moisture all the way to the Pacific. The tongue of low-salinity water that extends westward into the Pacific Ocean from Central America (**Fig. 7-15**) is evidence of this transport.

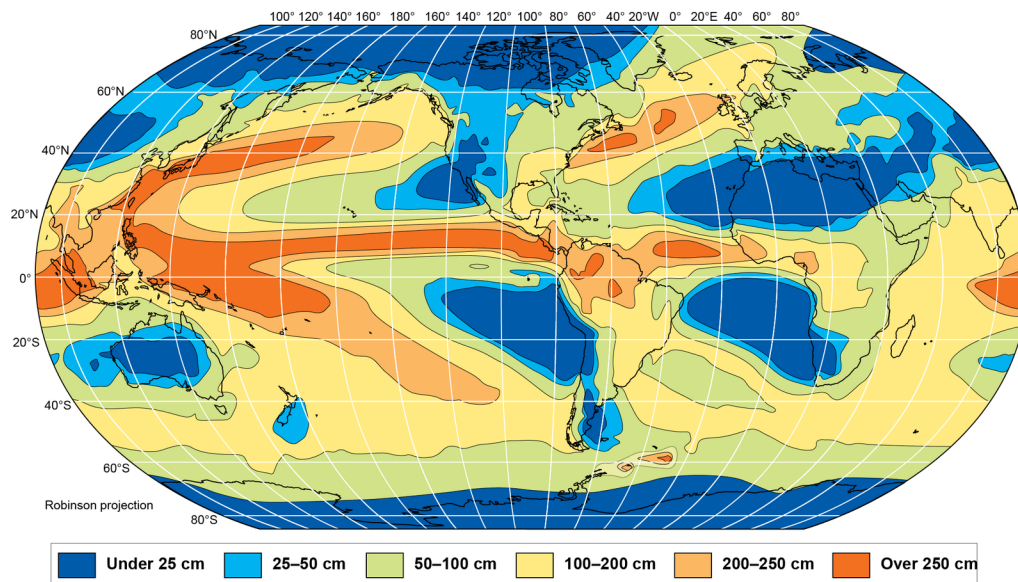


**Figure 7-17.** Distribution of mean annual atmospheric precipitation. Precipitation is high in the tropical atmospheric upwelling zone at the equator, generally low at mid latitudes in the atmospheric downwelling zones between the Hadley and Ferrel cells, higher in the atmospheric upwelling zones between the Ferrel and polar cells, and low again in the polar atmospheric downwelling zones. This pattern is more evident in the Southern Hemisphere because it is less modified by the presence of landmasses than it is in the Northern Hemisphere.

The Southern Hemisphere trade winds in the Atlantic Ocean carry their moisture predominantly into the Amazon and Orinoco river basins. Both of these major rivers discharge into the Atlantic Ocean equatorial region, where salinity is generally low.

Extremely high and extremely low salinities are most common in **marginal seas**, which have limited water exchange with the open oceans. Hence, if evaporation exceeds precipitation, salinity increases in such seas, and the high-salinity water is not mixed effectively with the lower-salinity water of the open ocean. Surface water salinity is high in the Mediterranean and the Arabian Gulf, and it is especially high in the Red Sea. All of these regional seas are in arid regions ([Fig. 7-17](#)) with low river runoff and high evaporation rates. Similarly, if precipitation and river runoff exceed evaporation, the salinity of a marginal sea is low. The best examples are the Baltic Sea and the marginal seas of Southeast Asia. In the Baltic, evaporation is low, and rainfall and runoff are moderately high ([Fig. 7-17](#)). In the marginal seas of Southeast Asia, rivers fed by monsoon rains discharge huge volumes of freshwater, and high rainfall rates overcome a relatively high rate of evaporation.

## 7.6: Interannual Climate Variations

In addition to seasonal variations, there are year-to-year and multiyear oscillations of climate that are related to ocean-atmosphere interactions. Oscillations have been identified in the tropical Pacific Ocean, the North Pacific Ocean, the Indian Ocean, the North Atlantic Ocean, and the Arctic Ocean. These may be linked with each other in ways we do not yet understand. One phase of the oscillation in the tropical Pacific Ocean, called El Niño, is associated with climate changes affecting much of the globe, and this oscillation has been extensively studied.

### El Niño and the Southern Oscillation

El Niño (Spanish for “the boy”) was named because it was first observed around Christmas time off the coast of Peru. The coastal and equatorial waters off Peru are normally regions where cold water from below the ocean’s surface layers is upwelled. [Chapter 8](#) explains how winds cause this upwelling. The upwelled water contains high concentrations of **nutrients** that are essential to phytoplankton growth ([Chap. 12](#)). Upwelling areas therefore have high primary productivity rates, and massive **schools** of anchovy feed on the abundant phytoplankton that grow in the Peruvian upwelling region ([Chaps. 12, 13](#)). Until they were **overfished** ([Chap. 16](#)), the anchovy populations were so large that they yielded about 20% of the total world fish catch.

### Symptoms of El Niño

The first sign of El Niño on the Peruvian coast is a temperature change in ocean surface water. Much warmer water (5°C or more warmer) displaces the normally cold surface water ([Fig. 7-18](#)). The less dense warmer water causes stable stratification of the water column and inhibits upwelling ([Chap. 8, CC1](#)), thus reducing phytoplankton populations that were nourished by the cold upwelled water. As the phytoplankton decline, the anchovy population that feeds on them plummets, and the fishery collapses. El Niño recurs generally every 3 to 5 years and usually lasts several months until the situation reverts to neutral. El Niño events have been documented in historical records as far back as 1726.

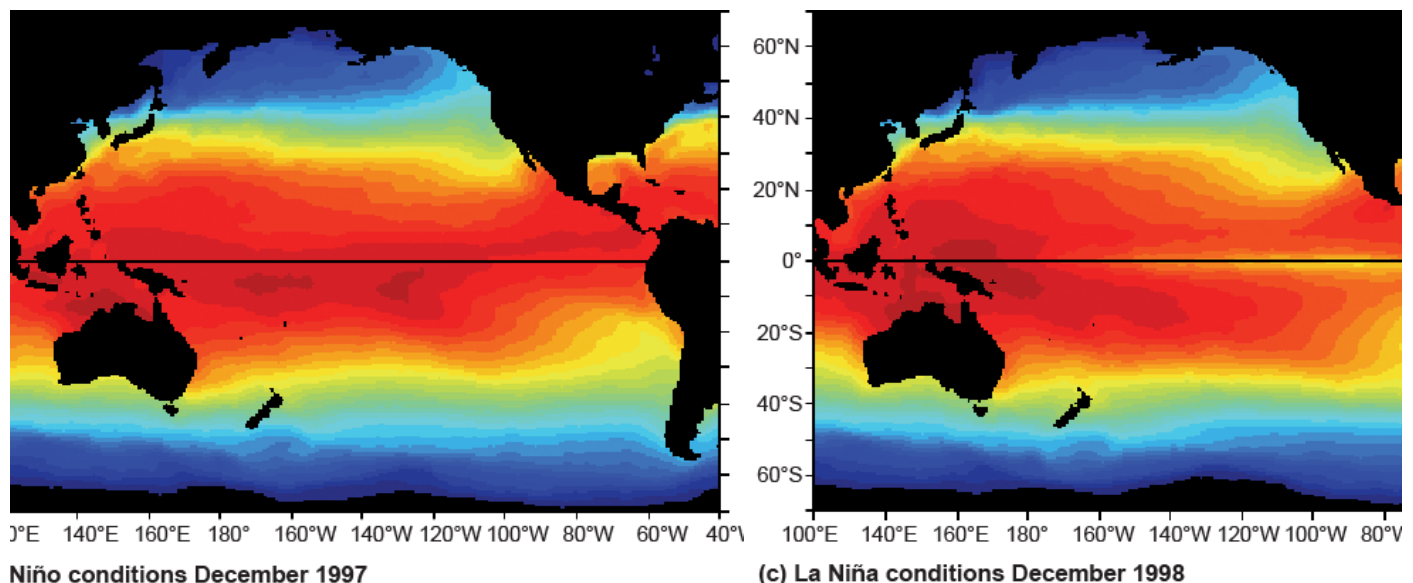


**Figure 18.** Sea surface temperature in the Pacific Ocean, as measured by satellite-mounted sensors. (a) During non-El Niño (neutral) conditions, there is a strong band of upwelling along the equator and in the Peruvian coastal zone, which can be seen in this December 1993 image by the yellow color that indicates lower surface water temperature in these areas. (b) During a strong El Niño (December 1997), there is no upwelling along the equator, and the Peruvian coastal upwelling is greatly reduced. (c) During La Niña conditions (December 1998), the upwelling area along the equator is substantially larger, and ocean water temperature around Indonesia is higher than normal because warm surface water is driven to the west.

El Niño is not an isolated event that occurs only off the coast of Peru. In fact, it is a complex series of cyclical changes in both ocean and atmosphere that affects much of the Earth’s surface. The series of changes has been called the “El Niño/Southern

Oscillation" (**ENSO**). The name reflects the fact that the changes in ocean and atmosphere involve a very large region of the Pacific, especially the region immediately south of the equator. The sequence of ocean-atmosphere interactions during an ENSO, now relatively well known, is described in the next section.

Between El Niños, a broad band of upwelling extends across the equatorial Pacific Ocean, and strong coastal upwelling occurs off the coast of Peru. The upwelling regions and their causes are discussed in [Chapter 8](#). The areas of upwelling are shown by colder surface waters ([Fig. 7-18a](#)). Trade winds move warm surface water to the west, where it accumulates near Indonesia ([Fig. 7-18a](#)). An atmospheric low-pressure area is well developed in this region because of the strong evaporation from the ocean surface and the high water temperature.

The low-pressure zone normally brings plentiful rains to Indonesia and the surrounding region. In contrast, a well-developed high-pressure zone off Peru maintains low rainfall over that country, which normally has an arid climate.

Trade winds push warm water to the west, and this warm water accumulates near Indonesia, elevating sea level by as much as a few tens of centimeters in extreme cases. For reasons not yet fully understood, this situation generally begins to change from November through April. The Indonesian atmospheric low-pressure zone and the Peruvian high-pressure zone both weaken and, in some years, actually reverse. During such periods, trade winds diminish and may reverse direction for several days. Trade winds do weaken in most years and return to normal by April, but in some years the weakening trade winds are followed by an El Niño.

## El Niño/Southern Oscillation Sequence

ENSO follows a relatively well-known sequence that occurs after the Indonesian low-pressure and Peruvian high-pressure systems weaken. The sequence begins as the strength of the southeast trade winds lessens in response to the weakening Peruvian high-pressure zone. The southeast trade winds are the cause of the Peruvian upwelling ([Chap. 8](#)). Therefore, the upwelling is reduced or may be stopped if the trade winds abate entirely or reverse direction.

Under neutral conditions, before an El Niño is initiated, the sea surface near Indonesia is elevated (a few tens of centimeters) in relation to the region near Peru ([Fig. 7-19c](#)). The difference is normally partially compensated for (the sea surface slope is balanced) by higher atmospheric pressure near Peru ([Chap. 8](#), [CC13](#)). However, the atmospheric pressure difference depends on the relative strength of the Indonesian low-pressure and Peruvian high-pressure zones. Because both the high- and low-pressure zones are weakened at this stage of ENSO, the horizontal pressure change within the tropical Pacific atmosphere is too small to balance the sea surface slope and surface water flows toward the east to lower the slope.

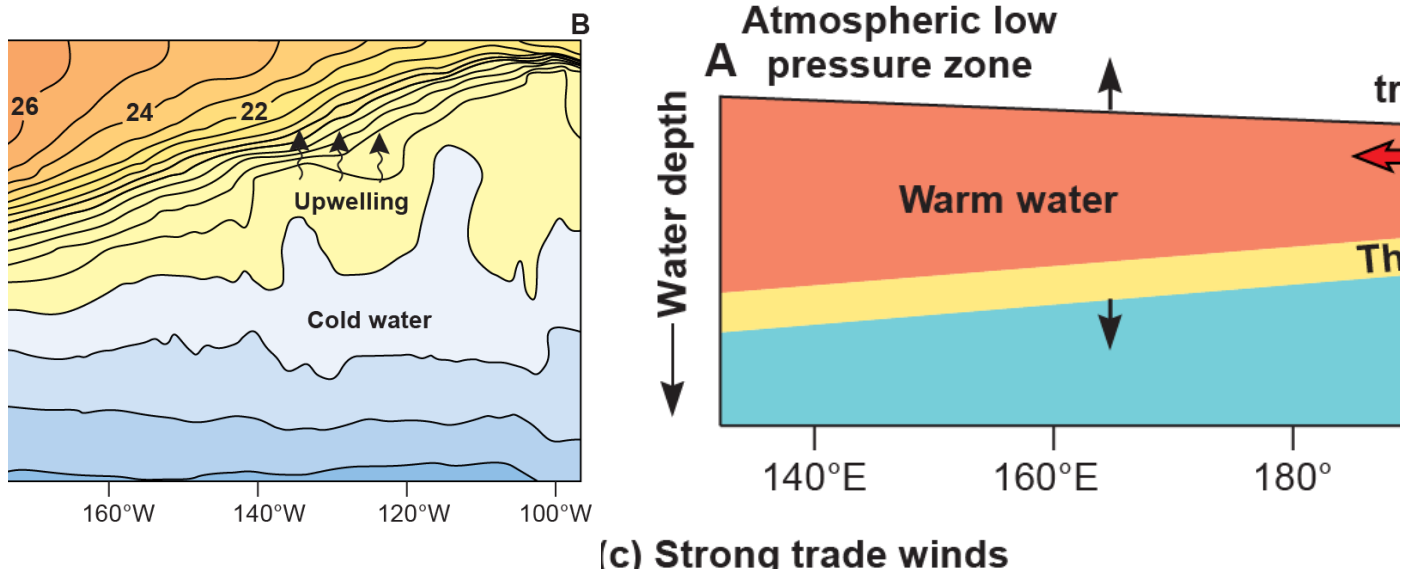


**Figure 7-19.** One possible model for the development of El Niño. (a) The area of the Pacific Ocean where El Niño develops. (b) When trade winds are strong, during neutral and La Niña phases, warm surface water is pushed westward and piles up on the western boundary of the Pacific Ocean near Indonesia. The surface layer is thinned near Peru, and upwelling occurs. (c,d) Vertically exaggerated cross sections show a simplified model of the interaction of trade winds with sea-level and thermocline depth in the two extreme modes of ENSO. The cross-section in part (c) shows the effect of strong trade winds on the sea surface and surface water layer. This is the normal, or in extreme cases, the La Niña phase. (d) When the trade winds diminish, surface layer water is transported eastward toward Peru, resulting in the El Niño phase. During an El Niño, the surface layer thickens near Peru, and upwelling ceases.

In response, warm surface water from the western tropical Pacific flows eastward toward Peru (**Fig. 7-19d**). The flow has characteristics of an extremely long-wavelength wave that travels directly from west to east along the equator, where there is no Coriolis Effect (**CC12**). Therefore, the wave is free to flow across the entire ocean without deflection. This is one of the unique characteristics of ENSO. The warm surface water flows eastward across the Pacific Ocean over the cold upwelled water that is normally found in this region (**Fig. 7-18b**). The warmer, less dense surface water arrives near Peru and establishes a steep **pycnocline** in December, about 9 months after the event has started. Because steep pycnoclines are effective barriers to vertical mixing and water movements, the Peruvian coastal and tropical Pacific upwelling is further inhibited.

El Niño usually persists for about 3 months, but in extreme cases it may last 15 months or more. It ends when the reestablished trade winds again begin to drive the warm surface water of the tropical region to the west, thus restarting the upwelling process near Peru.

The oscillation set in motion by El Niño sometimes appears to “overshoot” as the system recovers. The trade winds become stronger, the water near Indonesia becomes warmer, and upwelling is stronger and water temperature lower near Peru (**Fig. 7-18c**). In addition, the low-pressure zone near Indonesia and the high-pressure zone near Peru become especially well developed. This situation is known as La Niña (“the girl”), in contrast to El Niño (“the boy”).

## Effects of El Niño

The climatic effects of El Niño are felt far beyond the tropical Pacific. It appears that the extended area of warmer-than-usual tropical waters in the Pacific enhances the transfer of heat energy toward the poles. This change is manifested in many ways as the atmospheric convection cell system responds to the stimulus. El Niño differs in strength and other characteristics that make each El Niño’s effects somewhat different, but the same general pattern of effects is always present (**Fig. 7-20**).

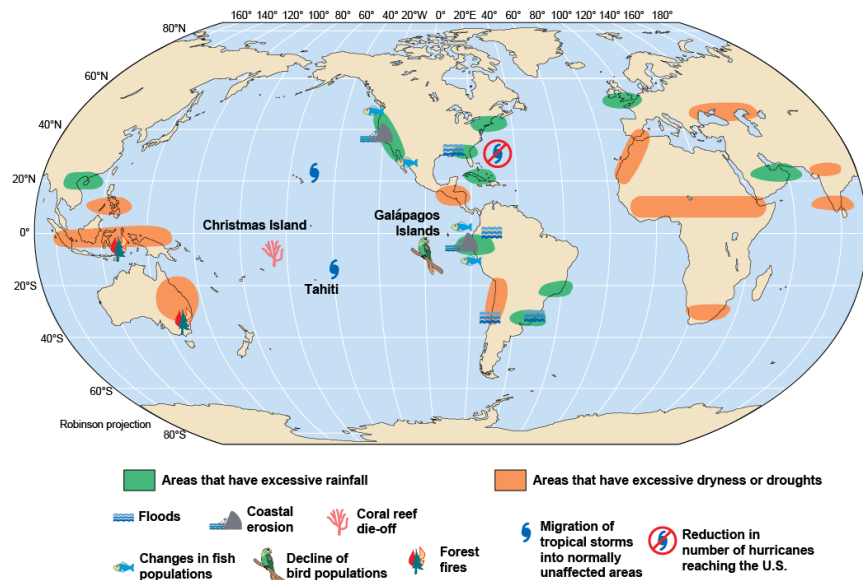


**Figure 7-20.** Strong El Niño events have worldwide effects that, in different areas, include droughts, excessive rainfall and floods, coastal erosion, tropical storms in areas where they do not normally occur, and various changes in terrestrial and marine ecosystems, some of which can be disastrous.

In the tropical Pacific, El Niño's effects are sometimes devastating. Reduction of the warm-water pool and weakening of the atmospheric low near Indonesia brings droughts to this normally high-rainfall region. In contrast, Peru has heavy rains and coastal flooding as sea level rises. The sea level rises because the surface water layer is warmer and less dense than normal. Consequently, the surface layer is "thicker" than the normally higher-density and cold surface water layer. Offshore from Peru, the marine **food web** collapses from lack of nutrients normally supplied by upwelled waters. In extreme cases, the result can be massive die-offs of marine organisms, which decay, strip oxygen from the water, and produce foul-smelling sulfides ([Chap. 12](#)). The principal fishery, anchovy, is not the only biological resource affected. In severe El Niños, lack of food decimates the huge colonies of seabirds that live on islands near Peru, and causes penguins and **marine mammals** to undertake unusual migrations and probably suffer population losses.

Changes in atmospheric circulation caused by El Niño also alter ocean water temperatures in areas far from the tropical Pacific. In the particularly strong 1982–1983 El Niño, subtropical fishes were caught as far north as the Gulf of Alaska. This El Niño caused droughts in Australia, India, Indonesia, Central America, west-central South America, Africa, and Central Europe. Excessive rains in many cases plagued parts of Southeast Asia, California, the east coast of the U.S., and parts of South America, Britain, France, and the Arabian Peninsula ([Fig. 7-20](#)). La Niña, in contrast, typically brings flooding rains to India, Thailand, and Indonesia, and drought to Peru and northern Chile.

## Modeling El Niño

We have records of El Niños occurring regularly for centuries. La Niñas also may have occurred regularly, but they have less dramatic effects and were not recorded until scientific studies began. However, there is considerable variability from one decade to the next. In some decades, the cycle was relatively inactive; in others, it was pronounced. Since 1950, El Niño has occurred quite regularly and was followed regularly by La Niña throughout most of the period ([Fig. 7-21](#)). During the twentieth century, the strongest El Niño occurred in 1982–83 and two consecutive El Niños occurred without an intervening La Niña several times. In the first quarter of the 21st century, there were 7 El Niño episodes. Four of these El Niños were strong, and the 2015–16 El Niño almost tied the 1997–98 episode's record for the strongest El Niño in recorded history. Because El Niño can have worldwide impacts on ecosystems and fisheries, and can cause widespread droughts and floods, major scientific efforts have been made to develop mathematical models ([CC10](#)) that can predict the occurrence of the phenomenon ahead of time. A number of models have been developed, each calibrated with historical data. As models have developed, they have become increasingly sophisticated, requiring increasingly detailed atmospheric and oceanic data. Gathering these data is one of the most intensive and comprehensive observational programs ever undertaken in the Earth sciences.

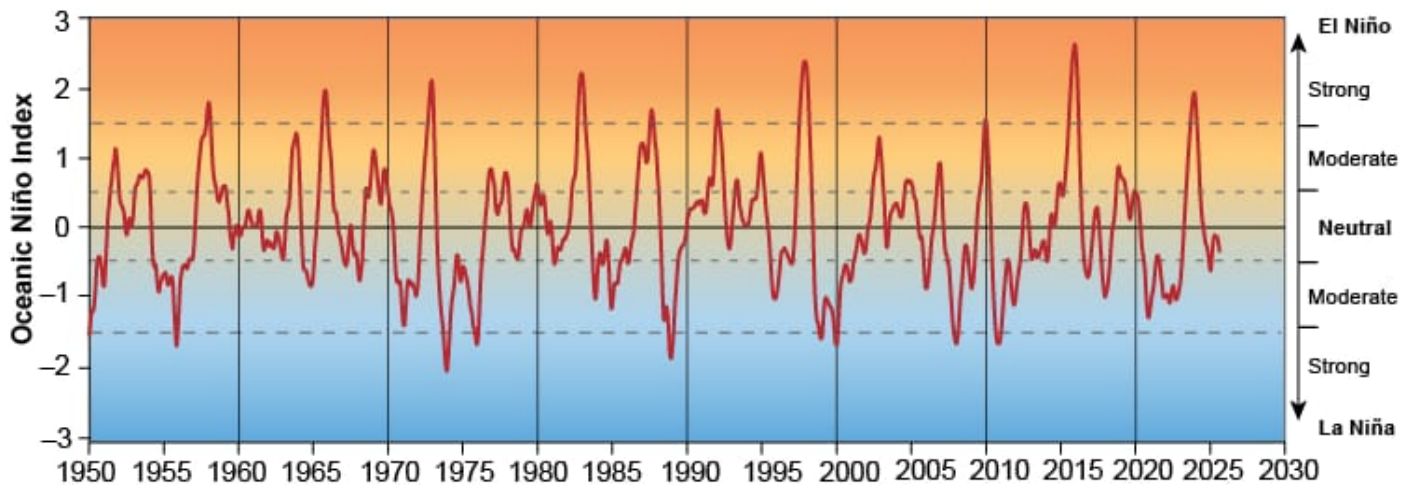


**Figure 7-21.** The NOAA Relative Oceanic Niño Index since 1950. The index is based on the average surface water temperature in a defined region of the central tropical Pacific Ocean compared to the surface water temperature of global tropical oceans. El Niño (when the index rises above 0.5) and La Niña (when the index falls below -0.5) have occurred regularly about every 2–7 years.

By 1997, several models were sufficiently developed to make predictions, and all of them forecast that a modest El Niño would occur in the winter of 1997–98. The model predictions were only partly correct, as the El Niño was not a modest one. It turned out to be the most intense event on record, costing an estimated 23,000 lives and \$33 billion (1998 dollars) in damage worldwide. What went wrong?

Fortunately, this El Niño was extensively observed by satellite sensors, moored instrument arrays, autonomous floats, research vessels, and aircraft. When the data were analyzed, the cause of the model failures appeared to be a burst of westerly winds in the western Pacific that occurred just as the eastward warm-water flow of the El Niño phenomenon was starting. These winds, unanticipated in the models, appear to have been timed perfectly to increase the flow of warm water across the ocean and, thus, intensify the El Niño.

Further research suggested that similar bursts of westerly winds also may have occurred just as other unusually intense El Niños had started. It has been hypothesized that these westerly winds are part of another ocean-atmosphere oscillation, called the Madden-Julian Oscillation (MJO). About every 30 days, this oscillation causes bursts of winds that are associated with an eastward-moving patch of tropical clouds. These clouds extend high into the troposphere and are easily observed from satellites. If this hypothesis is correct, the most intense El Niños may turn out to be formed only when the timing of this burst and the ENSO sequence are exactly right. The MJO cannot yet be reliably forecasted well in advance, and it may never be if the system is **chaotic (CC10, CC11)**, which is very likely. Unfortunately, this would mean that the intensity of an El Niño could not be accurately forecast until it had already started.

## El Niño Modoki

Since the early 1990s, scientists have noted a new type of El Niño that has been occurring with greater frequency and has been getting progressively stronger. In an El Niño Modoki (Japanese for “similar but different”), the maximum ocean warming is found in the central-equatorial, rather than eastern, Pacific. Such central Pacific El Niño events were observed in 1991–92, 1994–95, 2002–03, 2004–05, and 2009–10. Many of the global climate effects of an El Niño Modoki are similar to those that occur in a classic El Niño, but others are significantly different. For example, during a regular El Niño, there are generally fewer Atlantic hurricanes that make landfall in the U.S., whereas in an El Niño Modoki, there are more such hurricanes. Since detailed instrument and satellite studies of ENSO have covered only several decades, researchers are not yet sure whether El Niño Modoki is evidence of human-induced global climate change in the ENSO pattern or whether it is simply a natural long-term variability of ENSO. Some model studies have hypothesized that global warming due to human-produced greenhouse gases could shift the warming center of El Niños from the eastern to the central Pacific, further increasing the frequency of El Niño Modoki events in the future, but recent El Niños do not yet appear to show this trend.

## Other Oscillations

ENSO is the best-known ocean-atmosphere oscillation, and it is believed to be the most influential in causing the interannual

variations of the Earth's climate. A number of other oscillations are known, and they too have an influence on climate, but primarily just in specific regions. The MJO, discussed in the previous section, is one such oscillation, but there are several others that have longer periods and greater effects on regional climates. These include relatively well-known oscillations in the North Atlantic and North Pacific Oceans: the North Atlantic Oscillation (NAO) and the Pacific Decadal Oscillation (PDO), respectively. More recently, oscillations have been observed in the Arctic Ocean and in the Indian Ocean. The details of what is known about these oscillations and their effects on regional climates are beyond the scope of this text. However, the following paragraphs provide a very brief overview of each.

The North Atlantic Oscillation is a periodic change in the relative strengths of the atmospheric low-pressure region centered near Iceland and the subtropical high-pressure zone near the Azores. This pressure difference drives the wintertime winds and storms that cross from west to east across the North Atlantic. When the gradient between the two zones weakens, more powerful storms create harsher winters in Europe. The shift can affect marine and terrestrial ecology, food production, energy consumption, and other economic factors in the regions surrounding the North Atlantic. The periodicity of NAO is highly variable, but it ranges from several years to a decade or more. The Arctic Oscillation is closely coupled with the NAO. The Arctic high-pressure region tends to be weak when the Icelandic low is strong. The strength of the Arctic high affects climate around the Arctic Ocean, as well as the direction and intensity of ice drift.

Two modes characterize the Pacific Decadal Oscillation. In one mode, the sea surface temperatures in the northwestern Pacific, extending from Japan to the Gulf of Alaska, are relatively warm, and the sea surface temperatures in the region from Canada to California and Hawaii are relatively cool. In the other mode, these relative temperatures are reversed. In the first mode, the jet stream flows high across the Pacific and dips south over the Pacific coast of North America. Pacific storms follow the jet stream onto the continent over Washington, Oregon, or British Columbia, and California is cool and dry. Moist, relatively warm air is transported across the northern half of the United States, resulting in relatively mild, but often wet, winters in this region and in states influenced by warm, moist air masses from the Gulf of Mexico. In the second mode, the jet stream and Pacific storm tracks parallel the Pacific coast into Canada and Alaska. Parts of Alaska and much of California experience a mild winter. However, the jet stream's location allows cold, dry Arctic air to flow over the central United States, pushing the Gulf of Mexico air mass south, and a generally cold winter occurs in the eastern United States. We discuss some of the effects of the PDO in more detail in the next section.

In the Indian Ocean, there is no strong oscillation comparable to ENSO in the Pacific. However, there is evidence of a weaker oscillation that is similar. Although the Indian Ocean oscillation, known as the Indian Ocean Dipole (IOD), is much weaker than ENSO, it exhibits similar warm, neutral, and cold phases. The warm phase is characterized by warmer water on the western side of the ocean and cooler water on the eastern side, while the cold phase has warmer water in the eastern Indian Ocean and cooler water in the west. Modeling has shown that the intense drought of 1998–2002 over a swath of mid latitudes spanning the United States, the Mediterranean, southern Europe, and Southwest and Central Asia was possibly caused by a combination of the strong El Niño of 1997–98, and the cool phase of the IOD oscillation in the Indian Ocean.

## Climate Chaos?

The decade following the 1976–77 winter was extremely unusual in the U.S. and elsewhere. For example, in the following eight years, five severe freezes damaged Florida orange groves. Such freezes had occurred, on average, only once every 10 years over the previous 75 years. In addition, waves 6 m or higher tore into the southern California coastline 10 times in the 4 years from 1980 to 1984. Such waves had occurred there only 8 times in the preceding 80 years.

Were all these “unusual” events the result of a sudden shift in climate or just due to normal year-to-year variations of the Earth's climate? Because any one such change could be just a year-to-year variation, researchers examined the records of 40 different environmental variables that reflect climatic conditions around the Pacific region. The variables include wind speeds in the subtropical North Pacific Ocean, the concentration of **chlorophyll** in the central North Pacific Ocean, the salmon catch in Alaska, the sea surface temperature in the northeastern Pacific, and the number of Canada goose nests on the Columbia River. The 40 variables were combined to derive a single statistical index of climate for each year from 1968 to 1984.

The index showed a distinct and abrupt step-like change in value between 1976 and 1977, which is truly remarkable because such a clear signal of changed conditions is seldom, if ever, found even in any single environmental variable. Usually the many complex natural variations would obscure such a pattern. Was this change real? Did it indicate a chaotic “phase shift” between two relatively stable sets of environmental conditions? This study was perhaps the first evidence that chaotic shifts do indeed occur. It later became known that this shift was associated with the Pacific Decadal Oscillation (PDO), named in 1997. In 2000, this study was repeated, but this time using as many as 100 variables: 31 time series of atmospheric or oceanic physical variables, such as indices of the sea surface temperature; and 69 biological time series, such as catch rates of salmon. The analysis was

performed for the period 1965–1995, but because data were not available for all indices for each year, the analysis was done in two blocks—1965–1985 and 1985–1995—using a slightly different suite of indices for each. The results were as startling as those of the early studies. Distinct phase shifts were identified between 1976 and 1977, and between 1988 and 1989 (**Fig. 7-22**). These dates correspond with reversals of the PDO. Further evidence that these phase shifts may involve elements of chaos comes from the observation that some of the individual biological indices that changed in 1997 did not simply reverse in 1989. Indeed, some made step increases in both years, and some made step decreases in both years. These observations indicate that some species did not return to their former numbers in the ecosystem. This indicates that atmospheric oscillation driven shifts may be an agent of long-term change in the success of individual species within ocean and, perhaps, terrestrial ecosystems.

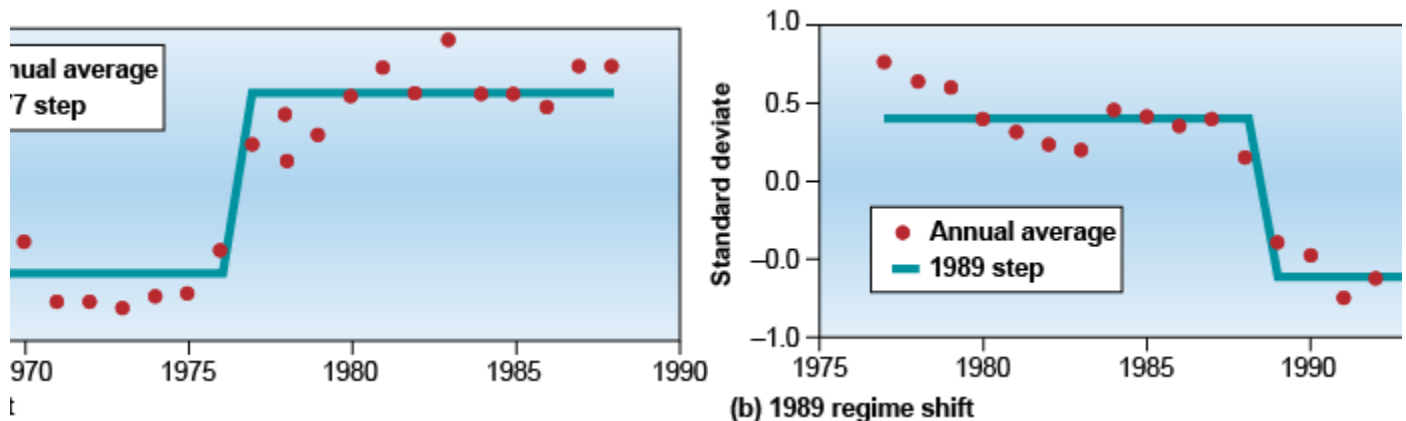


**Figure 7-22.** Variations in a composite index of environmental variables related to climate in the North Pacific Ocean. Time series were examined for 100 variables—31 climatic and 69 biological—although data were not available for all variables throughout the analysis period. (a) A regime shift occurred in the winter of 1976–1977, as shown by the step change in this function, which includes time series for 91 variables. (b) A regime shift also occurred in 1989, as shown by the step decrease in an index that includes time series for 94 variables. However, this second shift did not restore the system to its pre-1977 state.

If changes in the Earth's ocean-atmosphere climate system and associated ecological effects are chaotic, what does this mean? Chaotic systems often remain stable, varying within a well-defined range of behaviors, until the entire system suddenly shifts to a new stable state and varies within a different range of behaviors (**CC11**). Thus, it may mean that ecosystems are constantly changing in unpredictable ways. It may also mean that, even if no dramatic climate changes have yet occurred in response to human releases of greenhouse gases, there is no guarantee that any future changes will be small or slow. Drastic, sudden climate changes could be triggered if greenhouse gas concentrations in the atmosphere continue to increase, or they may already be inevitable. Decades of research on the Earth's ocean-atmosphere climate system are needed before we can have confidence in any forecasts of future climate in a world in which we have caused an unprecedented, rapid increase in the concentration of greenhouse gases in the atmosphere. Furthermore, such forecasts may never be more reliable than the local weather forecast.

## 7.7: Land-Ocean-Atmosphere Interactions

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Landmasses interact with atmospheric circulation in two ways. First, they can physically block or steer air masses moving across the Earth's surface. Air must rise over mountain chains or flow around them. Second, the thermal properties of landmasses and of the ocean are very different because rocks and soil have lower heat capacities than water has. Consequently, land warms more quickly than the oceans under the same solar radiation intensity. Also, land cools and gives off heat to the atmosphere much more quickly than the oceans do.

Interactions of landmasses with the atmosphere and oceans exert substantial control over the climate characteristics of most of the Earth's land and much of its oceans. The effects of the interactions on climate may be extremely localized and limited to one small area, or they may range across an entire continent or ocean. On the local scale, effects include land and sea breezes and the "island effect," which are discussed later in this chapter. On a larger scale, effects include the monsoons and the climate characteristics of the interior of continents. In the following sections of this chapter we discuss how land-ocean-atmosphere interactions can occur on different geographic scales. We review first the global interactions that determine climate, then larger-scale weather events that extend or travel across large regions of a continent or ocean, and finally smaller-scale weather features that occur locally.

## 7.8: Global Climate Zones

Climate in a given region is defined by several factors, including the average annual air temperature, seasonal range of temperatures, range of temperatures between day and night, extent and persistence of cloud cover, and annual precipitation and its seasonal distribution. Each factor varies with latitude and location in relation to land, mountain chains, and ocean. The factors also vary from year to year. Hence, most climate zones are not separate and distinct, but merge into one another. The demarcation lines drawn between the zones in [Figure 7-23](#) are therefore only approximations.

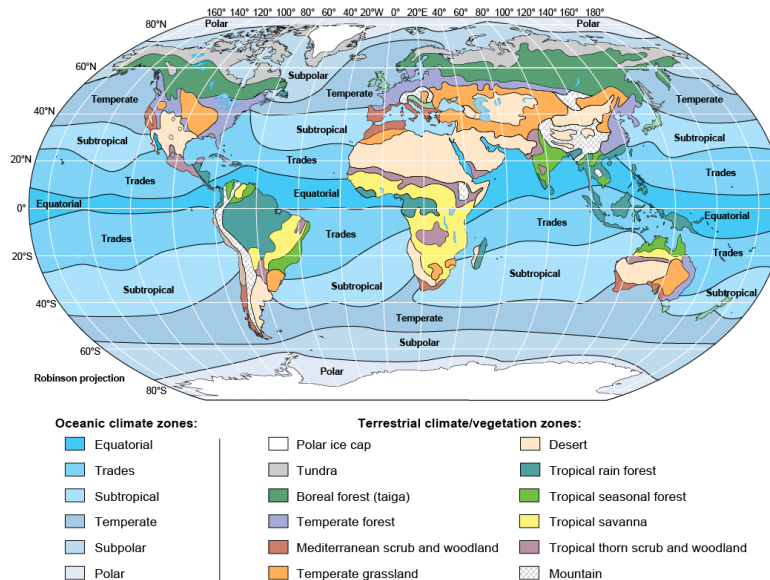


**Figure 7-23.** The climate zones of the oceans are arranged in bands that generally follow a latitudinal pattern and that are related primarily to ocean surface water temperature (see [Figure 7-14](#)). Terrestrial climate and vegetation zones follow a more complex pattern that reflects the presence of mountain chains, the prevailing direction of air mass movements, and the proximity to the oceans, all of which affect the distribution of rainfall and seasonal temperature ranges.

### Ocean Climate Zones

Climate zones over the oceans are generally latitudinal bands that parallel the average ocean surface temperatures shown in [Figure 7-14](#).

Polar ocean zones are covered with ice for much of the year, and ocean surface temperatures remain at or close to freezing all year ([Fig. 7-24a](#)) because of the heat-buffering effect of ice ([Chap. 5, CC5](#)). These zones have low precipitation and generally light winds (except where land and ocean interact). In subpolar zones, sea ice forms in winter but melts each year. Despite low precipitation amounts in the polar and subpolar zones, ocean surface salinity, particularly in the Arctic Ocean, is low. Surface water salinity is lowered during the continual freeze-and-thaw cycle that creates sea ice. As water freezes, **ice exclusion** causes dissolved salts to be excluded from the ice. The salts are left in the remaining water which raises the salinity and the resulting cold, high-salinity surface water sinks because its density is greater than the water below ([CC1](#)). When sea ice melts in the spring and summer, the non-salty water released from the melting ice mixes with the upper layer of ocean water and lowers its salinity. In the Arctic Ocean, freshwater runoff from the continents also contributes to low surface-water salinity.

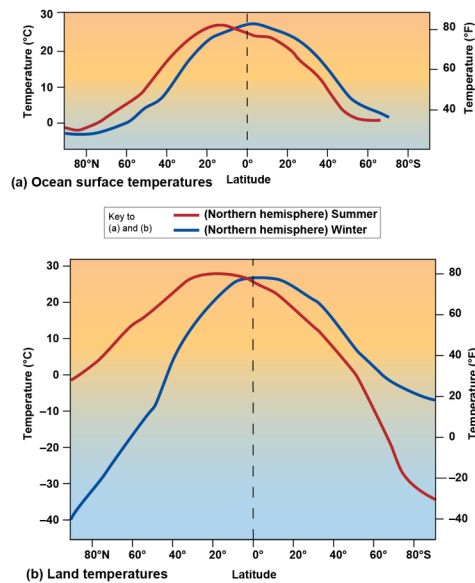


**Figure 7-24.** Seasonal and latitudinal variation in temperatures at the Earth's surface. (a) Average ocean surface water temperature. (b) Average temperature over land (or ice-covered ocean). Seasonal variation is much greater over land than over ocean, especially at high latitudes.

Temperate ocean zones are located in the zones of strong westerly winds (Fig. 7-11). They have high precipitation (Fig. 7-17) and are subject to strong storms called “extratropical cyclones.” These storms form especially in winter at the atmospheric polar fronts, which are locations where the polar and Ferrel cells meet (Fig. 7-10). The storms travel eastward and toward lower latitudes. Extratropical cyclones are described later in this chapter.

Subtropical ocean zones are located at the divergence between the atmospheric Hadley and Ferrel cells (Fig. 7-10). In these zones, winds are generally light, skies are usually clear, and precipitation is low. Many of the world's most desirable **beach** vacation areas are on coasts in these zones.

Trade wind zones are generally dry with limited cloud cover, but they are subject to persistent winds. Ocean surface water evaporation is very high, and thus, salinity tends to be high (Figs. 7-15, 7-16). In the equatorial region, ocean surface waters are warm (Fig. 7-14), evaporation is high (Fig. 7-16a), winds are light (Fig. 7-11), clouds are persistent, and rainfall is high and continuous throughout the year (Fig. 7-17).

## Land Climate Zones

Land (terrestrial) climate zones are not arranged into the same type of orderly latitudinal bands that characterize ocean climate zones (Fig. 7-23). The reasons for this difference include differences in thermal properties between land and water and the effects of mountains on air mass movements.

The average ocean surface temperature at any latitude changes little between seasons (Fig. 7-24a) because of the high heat capacity of water (Chap. 5, CC5). In contrast, soil and rocks have a much lower heat capacity. Consequently, seasonal changes in solar intensity cause substantial seasonal changes in average land-surface temperature (Fig. 7-24b). Seasonal changes are small in equatorial regions and increase with increasing latitude. The reason is that the annual range of daily total solar energy received at the Earth's surface increases with the distance from the equator.

Figure 7-24b shows the temperatures at continental locations far from oceanic influence. Daily and seasonal temperature changes are less in coastal locations than in regions far from the oceans at the same latitude (Fig. 7-25). The ocean provides a source of relatively warm air in winter and cool air in summer. Although the coastal air mass moderates climates in all coastal locations, the distance inland to which this effect reaches on a specific coast is determined by the prevailing wind direction and the location and height of mountain ranges.

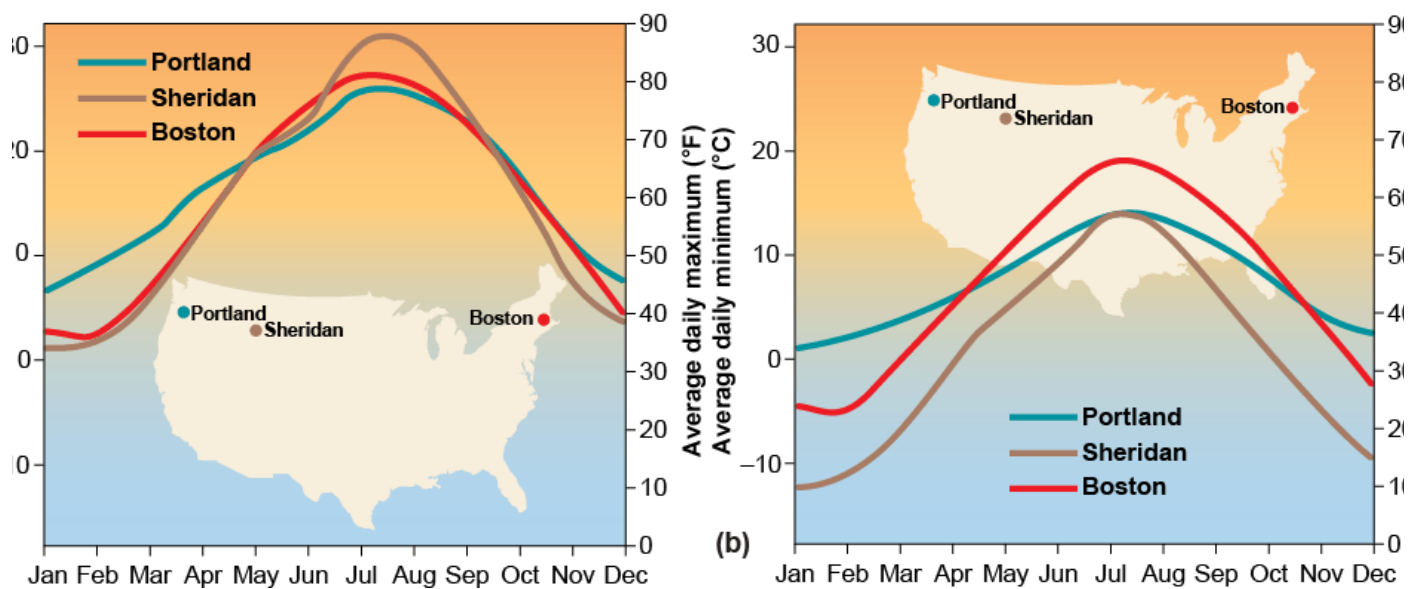


**Figure 7-25.** Annual variation in monthly average air temperature for Portland, Oregon; Sheridan, Wyoming; and Boston, Massachusetts. (a) Daily high temperatures. (b) Daily low temperatures. These cities are at approximately the same latitude; however, their climates are very different. Portland has a maritime climate with a small annual temperature range; Sheridan has a continental climate with a large annual temperature range; and Boston has a more continental climate than would be expected from its coastal location, because the climatic winds there blow from the continent toward the ocean.

If climatic winds blow onshore, coastal-ocean air can moderate temperatures and enhance precipitation many hundreds of miles beyond the coast. For example, warm coastal air that flows eastward across Europe is unimpeded by mountains. Hence, the moderate marine climate extends farther inland in western Europe than in the western parts of other continents, such as North and South America, where coastal mountains impede the westward flow of warm, moist air from the Pacific Ocean (Fig. 7-23). In the latter regions, the mountain effect (described in more detail later in this chapter) removes the principal source of the air mass's heat: its water vapor.

If climatic winds blow offshore, the moderating influence of the ocean is reduced. For example, Boston, Massachusetts, has a wider temperature range than Portland, Oregon (Fig. 7-25). Although both are coastal cities at about the same latitude, Boston is in a region of offshore climatic winds, while Portland is in a region of onshore climatic winds.

Coastal land climates are also affected by ocean currents. For example, Scandinavia and Alaska are at about the same latitude, but winters are much more severe in Alaska than in Scandinavia. In winter, most of Scandinavia is in the westerly wind zone and the temperate ocean climate zone, whereas most of Alaska is in the polar easterly wind zone and the subpolar ocean climate zone (Figs. 7-11, 7-23). In the North Atlantic Ocean, the Gulf Stream Current carries warm water from Florida north and west across the ocean to the seas around Scandinavia. The warm water provides heat energy through evaporation to moderate the Scandinavian atmosphere in winter. In fact, enough heat energy is provided to cause the westerly wind zone to extend into the region throughout the year (Fig. 7-11). A similar warm current, the Kuroshio Current, flows north and west from south of Japan toward Alaska (Chap. 8). However, the Aleutian Island arc deflects this warm current and it does not reach the Bering Sea on the west coast of Alaska.

The terrestrial climate zones shown in Figure 7-23 are defined primarily by seasonal temperature ranges and precipitation rates. In any given location, these parameters are controlled by the interaction of several factors, including ocean water temperatures, climatic winds, distance from the ocean, and the locations of mountain chains. A complete description of the zones is beyond the scope of this text, but the knowledge gained in this chapter can be used to explore the distribution of terrestrial climate zones depicted in Figure 7-23.

## 7.9: Weather Systems

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Wherever we live, we know from everyday experience that winds do not always blow in the directions of the climatic winds depicted in [Figure 7-11](#). Furthermore, contrary to the average climatic zone characteristics discussed previously, we know that some days are cloudy and rainy, and others are sunny and dry. These short-term variations, called “weather,” are caused by temporal and **spatial** variability in the motions of air in the atmosphere that can result from local variations in evaporation rate or in heating or cooling of the atmosphere. They can also result from **friction** as air flows over land or ocean surfaces. We can see this effect by observing how a smooth-flowing stream becomes churned as it interacts with rocks in rapids.

Atmospheric air movements are chaotic, and precise motions of any individual particle of air cannot be predicted (**CC11**). Therefore, we cannot ever hope to forecast weather precisely. Any forecast will be progressively less reliable the further ahead in time it is made. Weather forecasts made more than a few days ahead are unreliable and always will be. At best, future longer-range weather forecasts will be able to predict only trends in average weather that might occur over many days.

A complete discussion of weather systems is beyond the scope of this book. However, the behavior of high- and low-pressure zones, hurricanes, and extratropical cyclones is described briefly in the sections that follow.

### High- and Low-Pressure Zones

Atmospheric pressure at the Earth’s surface varies at any given location from day to day and over distances of tens or hundreds of kilometers across the surface at any given time. Once created, areas of slightly higher or lower pressure, which may cover hundreds or thousands of square kilometers, can persist for days or weeks. Atmospheric high- or low-pressure zones (“highs” and “lows” in weather media jargon) generate winds because air tends to flow on the pressure gradient outward from a high-pressure zone and inward toward a low-pressure zone. The moving air does not flow directly away from a high-pressure zone or toward a low-pressure zone, because the Coriolis Effect deflects the moving air mass (**CC12**).

The Coriolis Effect deflects moving air **cum sole** (which means “with the sun”)—that is, to the right in the Northern Hemisphere and to the left in the Southern Hemisphere. Thus, in the Northern Hemisphere, air that flows toward a low-pressure zone is deflected to the right. The deflection continues until the pressure gradient balances the Coriolis Effect. This balance results in a situation known as **geostrophic** flow, as discussed in **CC13**. A pattern is established in which winds flow counterclockwise around the low-pressure zone ([Fig. 7-26b](#)). Similarly, clockwise winds are established around a high-pressure zone ([Fig. 7-26a](#)). In the Southern Hemisphere, winds flow in the opposite direction: clockwise around a low-pressure zone and counterclockwise around a high-pressure zone. Friction between the moving air and the ground or ocean surface reduces the Coriolis deflection. Consequently, winds at, and near, the surface tend to spiral toward the center of low-pressure zones and away from the center of high-pressure zones.

Near-surface winds are influenced by friction and are therefore not true geostrophic winds, but they are in approximate geostrophic balance. As a result, their speed and direction can be reasonably estimated from isobaric charts that show the geographical variations of air pressure (**CC13**). Therefore, atmospheric pressure is among the most important measurements made at weather stations worldwide, and isobaric charts are a major tool for weather forecasting. **Wave height** forecasts also can be made from such charts because the relationships among wave height and wind speed, duration, and distance over which the wind blows (called **fetch**) are known ([Chap. 9](#)).

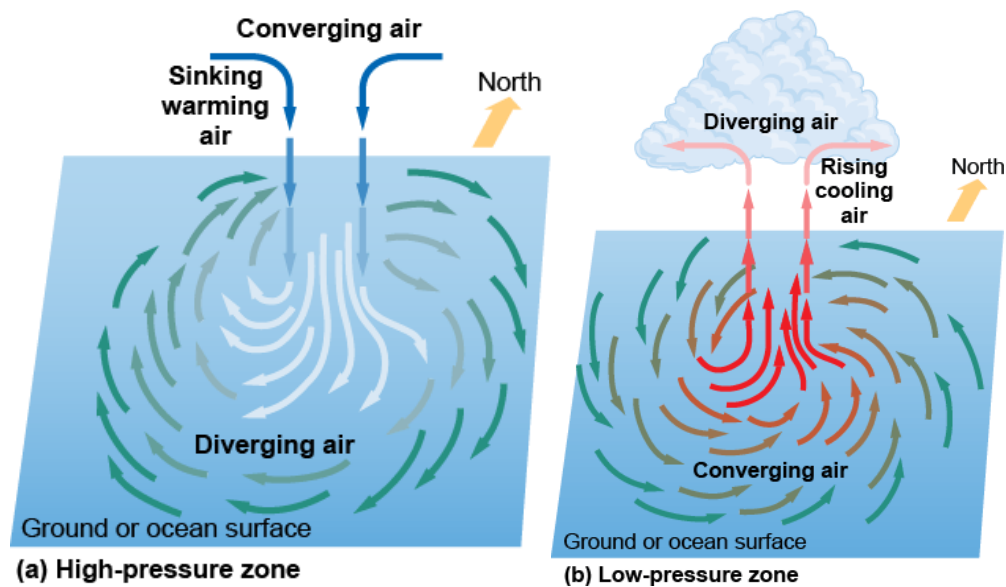


**Figure 7-26.** High- and low-pressure zones. (a) Cold, dense air sinks, creating a high-pressure zone in the lower atmosphere. The sinking air warms and flows out from the high-pressure zone. As air flows outward, it is deflected by the Coriolis Effect, forming winds that flow clockwise around the high-pressure zone (counterclockwise in the Southern Hemisphere). (b) Warm, moist air rises, creating a low-pressure zone in the lower atmosphere. The rising air cools, and water vapor condenses, causing precipitation. The warm rising air at the center of the low-pressure zone is replaced by air drawn toward the low from across the Earth's surface. As air flows inward, it is deflected by the Coriolis Effect, forming winds that flow counterclockwise around the low-pressure zone (clockwise in the Southern Hemisphere).

High- and low-pressure zones continuously form and re-form in the atmosphere and vary substantially in size and persistence. Low-pressure zones can be formed over the ocean where slightly higher ocean surface temperatures and/or higher evaporation rates occur. The result is an air mass in the low-pressure zone that has a temperature and water vapor content that is elevated compared to surrounding air. As increasing temperature and water content lead to a decrease in density, this air mass rises as a plume (**Fig. 7-26b**).

## Tropical Cyclones

**Tropical cyclones** are among the most destructive forces of nature. These intense storms form mostly during local summer or fall in both the Northern and Southern Hemispheres. Tropical cyclones develop over the warm ocean waters of tropical regions where a low-pressure zone develops that draws air inward across the ocean surface. If conditions are favorable, the low-pressure zone deepens (that is, the pressure drops further), and a tropical cyclone may form. In the Indian and western Pacific Oceans, tropical cyclones are called “cyclones” north of the equator and “typhoons” south of the equator. They are also called “baguios” around the Philippines and “willy willys” near Australia (**Fig. 7-29**). They are known as “hurricanes” only in the Atlantic and eastern Pacific Oceans, but they are the same phenomenon regardless of the name.

Winds in the developing low-pressure system are deflected by the Coriolis effect (**Fig. 7-26b**) until they flow around the low-pressure zone. Additionally, a centripetal (“center-seeking”) force helps confine motion to a curved path. As the air flows in toward and around the low-pressure zone across the ocean surface, it is warmed, gains moisture, and tends to rise. The winds thus blow toward the center, or “eye,” of the tropical cyclone in a rising helical pattern (**Fig. 7-27**). Because winds do not reach the eye to alleviate the low pressure, the low-pressure zone continues to intensify. The horizontal pressure gradient increases, and the wind speeds increase (**CC13**).

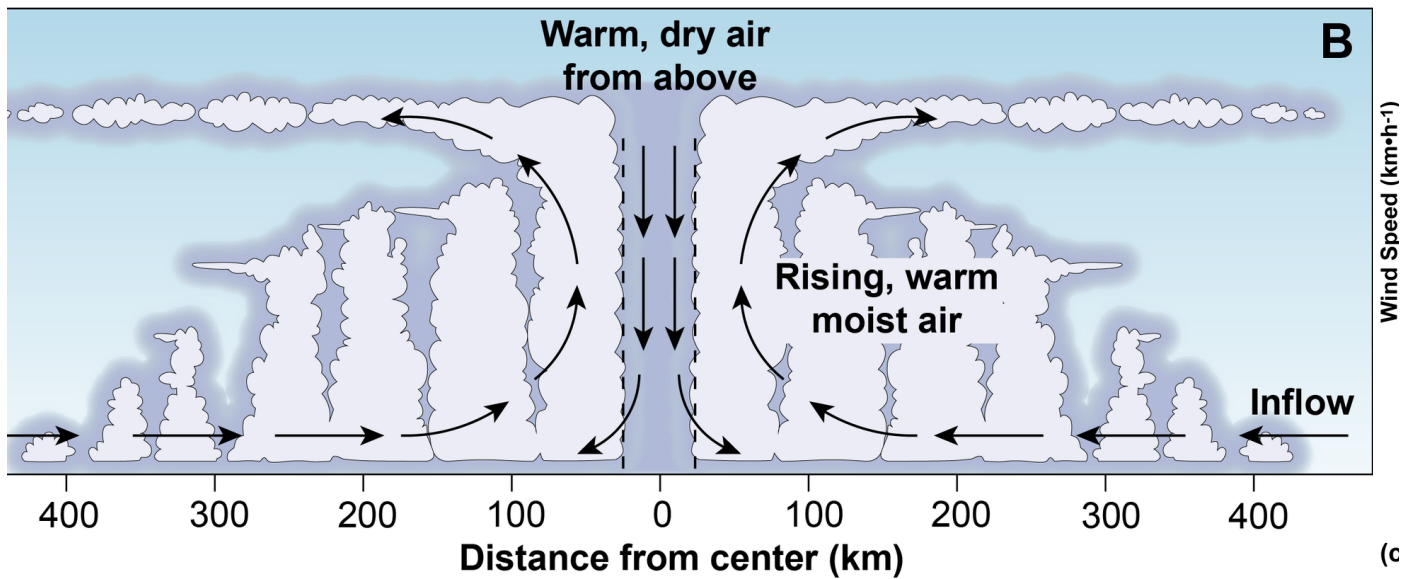


**Figure 7-27.** (a) Tropical cyclones are rotating storms in which air masses are drawn toward a low-pressure zone and rise in a helical pattern. This process forms spiral rain bands and a cloudless eye. (b) A vertical section of the hurricane across the line from A to B in (a) shows how the warm, wet air rises and spreads out at an altitude of more than 10,000 m. Dry air descends in the center of the storm and is heated adiabatically to form the eye. (c) A plot of wind speed and atmospheric pressure at the surface in the southwest-to-northeast cross section (A to B). Winds are strongest in the northeast quadrant of a hurricane, about 100 km from the eye of a typical storm, but the heaviest rainfall takes place closer to the eye. Storms may vary in size, and the maximum wind speed is higher when the atmospheric pressure in the eye is lower.

A well-formed tropical cyclone has an eye, usually about 20 to 25 km in diameter, within which winds are light and there are few clouds (**Fig. 7-27**). The eye forms when the low-pressure zone deepens enough to draw warm, relatively dry air downward into the eye, as winds are deflected from the eye in their rising helical pattern (**Fig. 7-27**). The eye is surrounded by a wall of clouds (the eye wall) where winds are moderate, but intense thunderstorms are created as the rotating air mass rises. Outside the eye wall is a region of intense winds that blow across the ocean surface (**Fig. 7-27**). The winds are counterclockwise in the Northern Hemisphere and clockwise in the Southern Hemisphere.

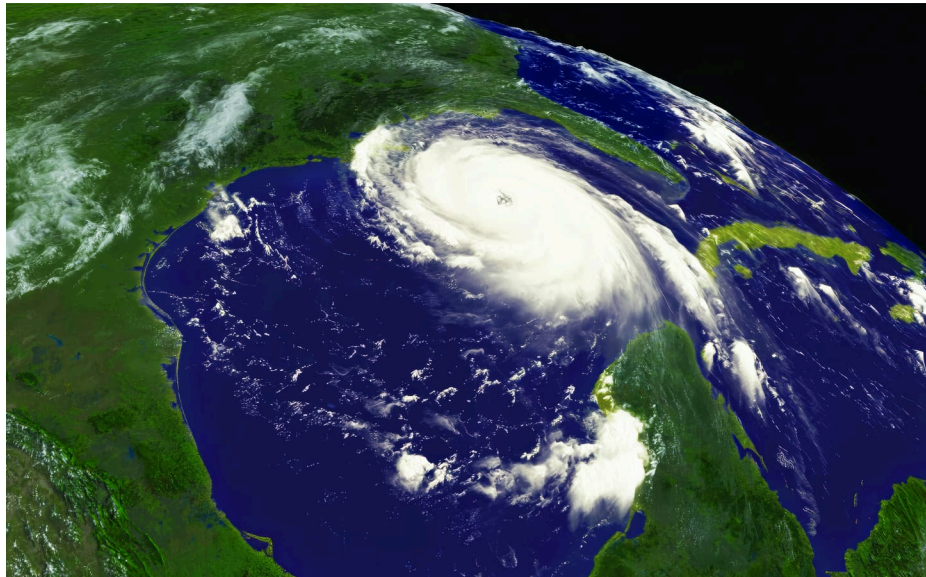


**Figure 7-28.** This satellite image of Hurricane Katrina (2005) was obtained as the hurricane moved across the Gulf of Mexico, just a few hours before it struck land, costing many lives and causing billions of dollars of damage in Louisiana and Mississippi. In the image, the outer bands of cloud are just reaching the coast near New Orleans while the trailing edge clouds extend as far as the Yucatan Peninsula in Mexico, so this monster storm was about 1000 km in diameter.

Tropical cyclones drift to the west in the prevailing trade wind direction at about 10 to 30 km·h<sup>-1</sup>, and sometimes faster. Their

westward paths are deflected by the Coriolis Effect, and they generally turn away from the equator. Over warm ocean water, hurricanes maintain or gain intensity as high temperatures and high winds encourage evaporation and feed more latent heat energy into the storm. When a hurricane moves over cold water or land, it quickly loses energy and releases its moisture as intense rainfall (10 to 20 billion tonnes of rainfall per day from an average hurricane).

Storm surge, freshwater flooding, wind, tornadoes, and surf and rip currents are all significant tropical cyclone hazards, some occurring far from the storm's landfall. The potentially most destructive and life-threatening effect of a hurricane has historically been the **storm surge**, or wave that pushes ahead of the storm. Storm surges cause flooding of low-lying coastal areas. However, improved early warning and evacuation procedures have reduced storm surge deaths, and drowning due to inland flooding and landslides caused by wind and rainfall from the tropical cyclone are now the largest cause of cyclone-related deaths in most areas.

The deadliest hurricane in U.S. history was the Galveston hurricane of 1900, that killed 8000 people, followed by the 1928 Okeechobee hurricane, that killed 2500. However, since 1961, satellite imagery and progressively improved forecasting have provided early warnings of approaching storms, which has dramatically reduced loss of lives.

In 1969, Hurricane Camille took more than 250 lives, and most of the victims drowned in the storm surge, which crested at nearly 7.5 m. Camille also caused almost \$1.5 billion in damage along the Gulf Coast. Hurricane Hugo, which hit South Carolina in 1988, and Hurricane Andrew, which hit South Florida in 1992, caused much greater damage than Camille but took few lives. Hurricane Andrew is estimated to have caused more than \$30 billion in damage. Hurricane Katrina came ashore in Louisiana and Mississippi in 2005, causing more than \$125 billion in damage and about 1500 deaths. In 2017, Hurricane Harvey hit Houston, Texas and killed at least 70, followed weeks later by Hurricane Irma that killed 134 as it traveled across Puerto Rico, Cuba, the Bahamas, and South Florida. This was the first year on record when two strong (category 4 on a scale of 1–5) hurricanes made landfall in the continental U.S., and the total damages are estimated to be as much as \$200 billion. Hurricane Harvey (2017) caused more than 100 deaths and Hurricane Helene (2024) caused over 250 deaths in the U.S. and both caused significant damage in inland areas. In the case of Helene, the majority of fatalities were caused by excessive rainfall well inland in the mountainous region of western North Carolina. The cost of damages in this region are estimated at almost \$80 billion.

Tropical cyclones form in all tropical oceans, but they are rare in the South Atlantic and eastern South Pacific ([Fig. 7-29](#)). Before March of 2004, only two tropical cyclones had ever been recorded in the South Atlantic Ocean, and neither had reached hurricane strength. However, in that month, Hurricane Catarina formed east of Brazil. The storm caused at least two deaths and substantial property damage when it came ashore on the southern coast of Brazil. Tropical cyclone damage and deaths are generally greater in the Islands of the western Pacific and mainland Asia than those in the U.S., as the frequency of these cyclones in the Pacific and Indian Oceans is higher than that in the Atlantic. The most vulnerable locations include the Philippines, which lie in the region with the highest frequency of these storms, and the densely populated low-lying areas of Bangladesh, India, Southern China, and mainland Southeast Asia, where evacuation of the large populations remains difficult and storm surge remains a major threat.

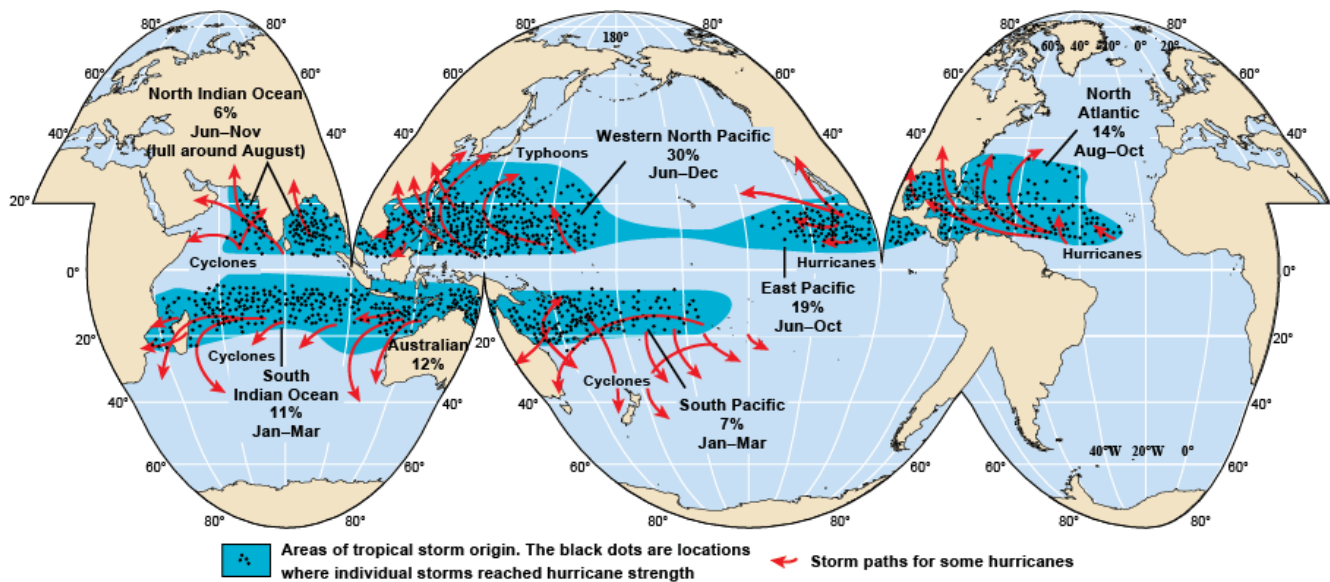


**Figure 7-29.** Tropical cyclones develop where the surface ocean water temperature is greater than 26°C, except in the equatorial region between 5°N and 5°S, where the Coriolis Effect is too weak to generate these storms. Surface waters are too cold for strong tropical storm formation in the eastern Atlantic Ocean and south of the equator in the eastern Pacific Ocean. Once formed, tropical cyclones travel generally east to west, curving toward higher latitudes as they are steered by the Coriolis Effect and by other weather systems. Tropical cyclones are most frequent in the western North Pacific Ocean, where they are called “typhoons” or “baguios.”

Tropical cyclones form only over surface waters with temperatures above about 26°C and do not generally form closer to the equator than 5°N or 5°S, because the Coriolis effect is too weak in this equatorial band.

Historical records suggest that the frequency and intensity of hurricanes may vary on a cycle lasting several decades. Climate models predict that hurricane frequency will not change significantly due to climate change in the next several decades, but they will become stronger on average, and the most powerful damaging storms will be more frequent and may maintain strength into progressively higher latitudes. For the U.S. and many other countries, this prediction is important. Coastal development has been intense over the past several decades, particularly in the southeastern U.S. and Gulf Coast communities within the hurricane band. In addition, sea level is rising, exposing more coastline to wave action. In the future, large coastal cities of the northeastern United States could be exposed to the full force of devastating hurricanes. Perhaps worse, future hurricanes might have much higher winds than the approximate maximum of  $200 \text{ km} \cdot \text{h}^{-1}$  observed in the most intense recent hurricanes. Studies of **sediment** deposited by ancient storms have provided evidence of “super” cyclones in Earth’s past with wind speeds probably higher than any ever directly observed and greater than what most buildings can withstand.

## Extratropical Cyclones

Powerful **cyclonic** storms form in the westerly wind regions in winter. Such storms form primarily at the polar front, where cold polar air moving toward the equator and the west and warmer mid-latitude air flowing east and toward the pole converge. Along the front between these two air masses, which are moving in opposite directions, waves periodically develop. The waves develop into cyclonic storms as the warm air flows over the top of the cold, denser air, and each moving air mass is deflected by the Coriolis Effect (**Fig. 7-30**).

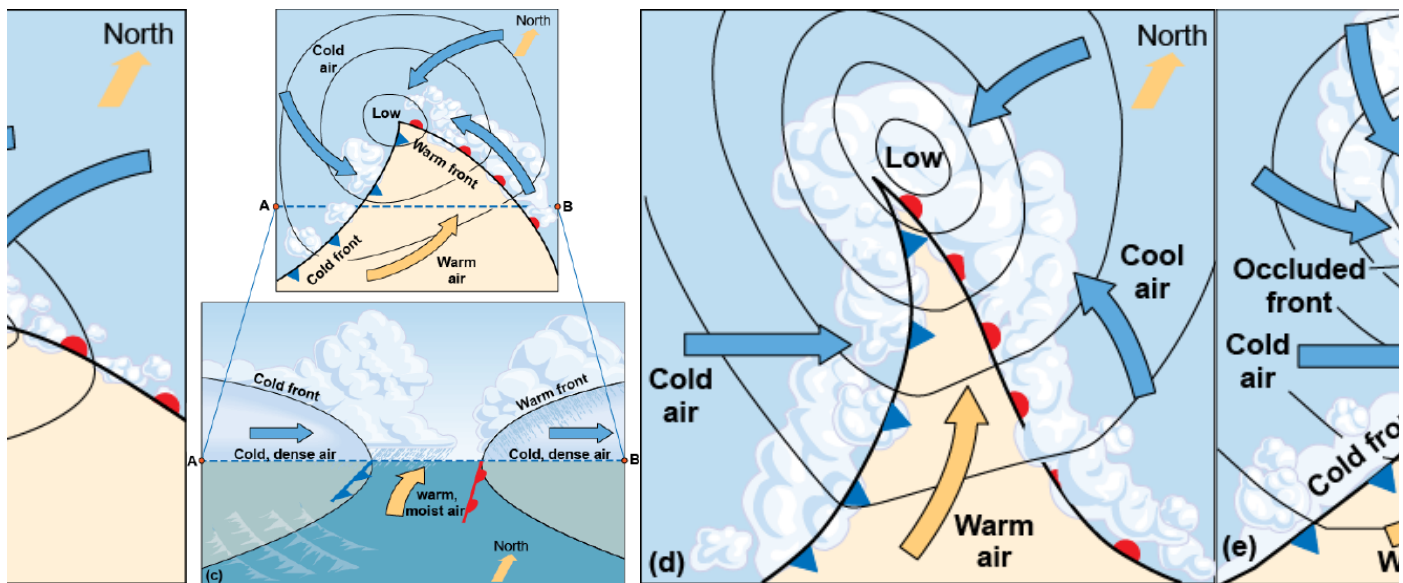


**Figure 7-30.** Formation of extratropical cyclones in the Northern Hemisphere. All the diagrams except (c) are surface-level map views. (a) The polar front separates cold air on the polar side of the front from warm air at lower latitudes. (b) Extratropical cyclones are formed when a wave develops on this front. (c) A low-pressure zone forms as warm, moist air flows north and rises, initiating counterclockwise rotation of the air masses around it (see Figure 7-26b). Once formed, the low-pressure zone may strengthen if the cold air mass flows over relatively warm ocean surface waters, is warmed, gains water vapor through evaporation, and also begins to rise. The vertical cross section shows that extensive areas of rain clouds form both where the advancing cold air mass forces warm, moist air to rise (cold front), and where warm, moist air is deflected to the right by the Coriolis Effect and rises over the cold air mass (warm front). (d) Eventually, the cold front catches up to the warm front as warm air continues to rise over the cold air mass. (e) This leads to the formation of an occluded front, which is the line where the warm and cold front meet. (f) The polar front reforms, and the extratropical cyclone slowly dissipates when warm air no longer flows into the central low-pressure zone, and atmospheric pressure rises. In the Southern Hemisphere, the rotation is in the opposite direction.

Storms that form at the polar front are called “extratropical cyclones.” They can be extremely large and carry huge amounts of moisture. They move eastward with the prevailing winds. Extratropical cyclones move more slowly than hurricanes and generally have somewhat lower, although still very high, wind speeds. They can be much larger than hurricanes and can cause storm surges. Consequently, they can cause severe and extensive wave **erosion** on exposed **shores**. The massive Pacific storms that bring winter rains, winds, and beach erosion to California and the Pacific Northwest are extratropical cyclones.

Extratropical cyclones can also intensify or deepen by interacting with a warm **western boundary current**. For example, the nor’easters that often strike the Mid-Atlantic coast of the United States in winter are extratropical cyclones that often undergo explosive intensification near the Gulf Stream after passing over the North American land mass.

## 7.10: Local Weather Effects

Interactions among ocean, atmosphere, and land are local in geographic scale. These interactions occur on or near the coast and sometimes produce large differences in weather and climate within a coastal region.

### Land and Sea Breezes

All parts of the Earth's surface are subjected to a daily cycle of heating and cooling as night follows day. The ocean's surface water absorbs solar radiation during the day without a significant temperature change for three main reasons. First, water has a high heat capacity. Second, much of the solar energy penetrates beneath the water's surface before being absorbed, particularly in low-**turbidity** waters. Third, wind stirs the upper water column, mixing and distributing the heated water. Similarly, ocean surface water cools slowly at night. Consequently, even in locations where daytime solar radiation is intense, the temperature of ocean surface water varies by, at most, a few degrees between day and night. The temperature of the land surface, in contrast, can vary over a much larger range **diurnally**. Anyone who has spent time in the middle of a continent, particularly in the mountains, may have observed the rapid heating and cooling of the land. A rock that is too hot to touch in the middle of a sunny day will feel cold within a few hours after sunset that evening.

Heat is transferred continuously between the land and the atmosphere, and between the ocean and the atmosphere. As a result, and because of its low heat capacity, the air next to the Earth's surface tends to change temperature relatively quickly to match that of the land or ocean surface.

During daylight hours, coastal water and its overlying air mass warm only slowly, whereas the adjacent land and its overlying air mass warm much more rapidly. Where air temperature is higher over coastal land, the air over land becomes warmer and less dense, expands vertically, and thus has reduced surface pressure relative to the cooler marine air (**Fig. 7-31a**). Cooler air from over the ocean is drawn inland by the resulting horizontal atmospheric pressure gradient to replace the rising heated air. This process creates onshore winds called "sea breezes." The sea breeze blows until the solar radiation is reduced in intensity and the land has cooled to the same temperature as the coastal surface waters. Sea breezes gradually weaken near and after sunset as land cooling reduces the thermal contrast.

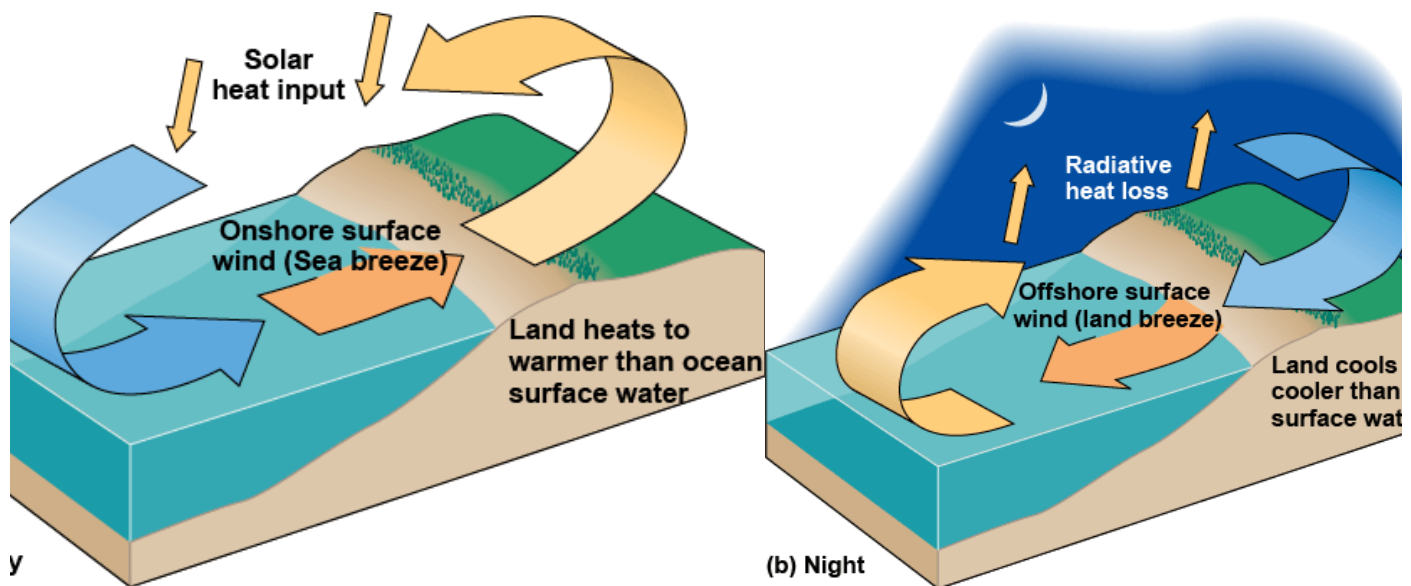


**Figure 7-31.** (a) Land heated by the sun during the day heats the ground-level air, which rises and is replaced by cooler air drawn in from the ocean. Thus, the surface wind blows from ocean to land and is called a "sea breeze." (b) At night, the land loses its heat quickly through radiation, but the ocean water remains at the same temperature. As the land surface cools to below the sea surface temperature, warmer air rises over the ocean and is replaced by cooler air drawn from over the land. Thus, the surface wind blows from land to ocean and is called a "land breeze."

As night progresses, the land and its overlying air mass cool rapidly until they are colder than the ocean surface water and its

overlying air mass (**Fig. 7-31b**). Air pressure increases over the land as the cooled, denser air mass sinks. This air mass now spreads toward the ocean, where atmospheric pressure is lower. Thus, a land breeze blows offshore during the latter part of the night and the early daylight hours (**Fig. 7-31b**).

Reversing local sea and land breezes are present on almost all coasts, but their intensity varies significantly with location, season, and larger-scale regional weather patterns. Probably for thousands of years, fishers in many countries have used sea breeze and land breeze patterns to their advantage, sailing offshore with the land breezes in the early morning hours and returning with the sea breezes late in the day. Many fishers still follow this pattern, even though they have motorized vessels.

## Coastal Fog

The famous fogs of San Francisco and other coastal locations are often related to sea breezes. In many coastal areas, especially where coastal upwelling occurs, surface waters near the coast are colder than those tens of kilometers offshore. When sea breezes develop, warm, moist air over the offshore water is drawn into the sea breeze system. As this air flows shoreward, it passes over cooler coastal water. If it cools sufficiently, some of the water vapor it carries condenses into extremely small water droplets, which we call “fog” (**Fig. 7-32**). Coastal fogs of this type occur along the entire West Coast of the U.S. and along the eastern coast from Maine into Canada. Similar fogs occur on the west coasts of South America and Africa and in many other locations.

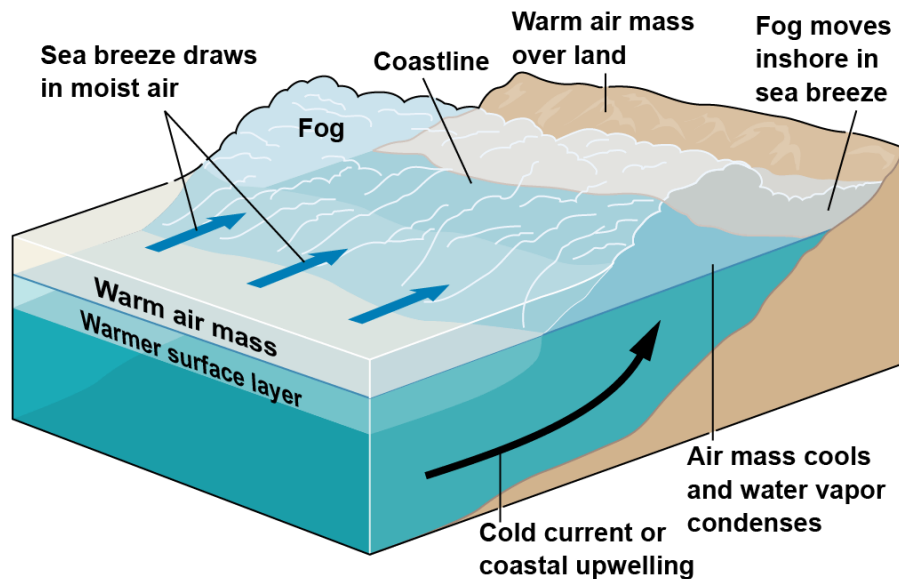


**Figure 7-32.** Coastal fog develops when moist air is drawn inland by a sea breeze or climatic winds and passes over a cold nearshore water mass. The cold nearshore water can be the result of cold-water currents or, more often, of coastal upwelling that brings cold, deep water to the surface.

## The Island and Mountain Effect

When air moving over the Earth’s surface encounters a mountain chain across its path, it is forced to rise up the mountain slope to pass the barrier. As the air mass rises, it cools by adiabatic expansion because the atmospheric pressure decreases with altitude. If the air is moist before it begins to rise and it rises high enough, water vapor will condense to form clouds, which eventually form rain or snow (**Fig. 7-33a**). Once the air has crossed the mountain ridge, it flows down the mountain slope. As it descends and pressure increases, the air warms, and the water vapor saturation pressure increases. Also, the air’s actual water vapor content remains low due to precipitation loss on the windward side. The leeward side of the mountains therefore, receives very little rain.

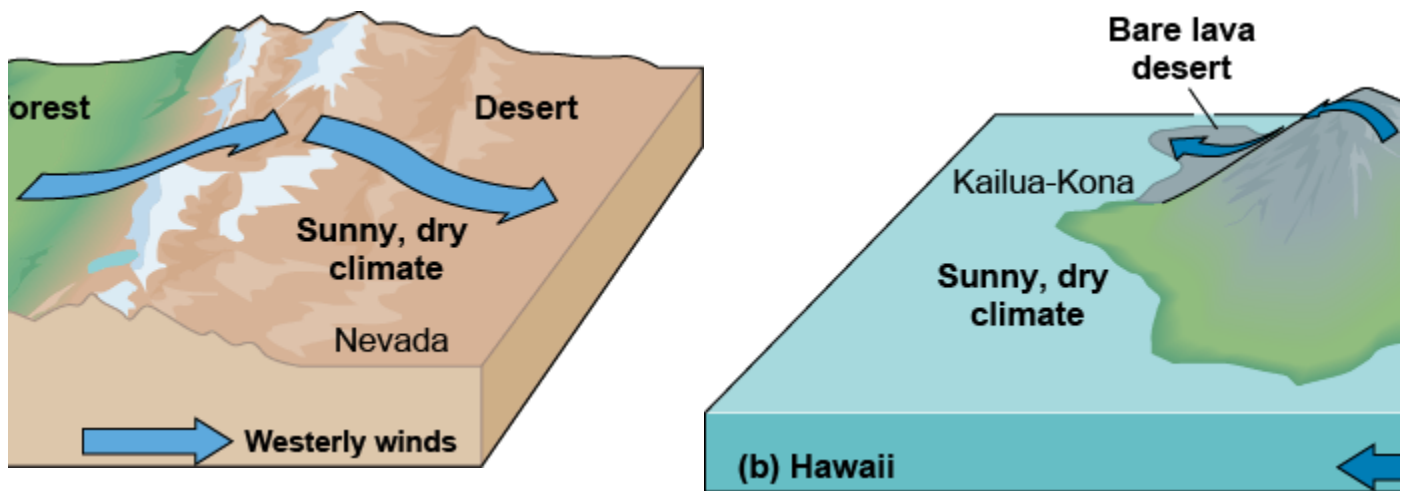


**Figure 7-33.** The mountain or island effect occurs where a moving mass of moist air, usually from over the ocean, encounters a mountain or mountainous island, rises, cools adiabatically, and loses water as rain or snow. Once over the mountains, the air mass descends, warms, and has a greater capacity for moisture. As a result, rain and snowfall stop. (a) The Sierra Nevada mountains of California intercept moist air moving inland from the ocean. Rain and snowfall are abundant on the California side of the mountains, but the Nevada side is a desert. (b) On the Hilo side of Hawaii, trade winds blowing from the northeast produce clouds, plentiful rainfall, and tropical rainforests, but the opposite side of the island, near Kailua-Kona, is sunny and has low rainfall.

In many coastal locations, the wind blows consistently from the ocean onto land. If a coastal or near-coastal mountain chain of sufficient height is present, the coastal side of the mountain range will have high rainfall, and the landward side will be desert. The best examples are in parts of South America, where the Andes Mountains divide the fertile coastal region of Chile from the desert or near-desert interior of Argentina, and in North America, where the west side of the Sierra Nevada chain is fertile and the east side is desert. Note that the coastal mountains of northern California are too low and have too many gaps to dry the air masses completely as they move off the Pacific toward the Sierra Nevada.

Skiers in the Sierras can use the mountain effect to decide where snow conditions are best. Resorts on the California side usually have more snow, particularly in poor snow years. However, the snow is generally wet and heavy. The resorts on the Nevada side, just over the ridge **crest**, generally get less snow. However, snow that does accumulate falls through dry, undersaturated air and often has a dry, powdery character that is desirable for skiing.

The mountain effect also operates on mountainous islands, where it is called the “island effect.” The island effect is extremely important to the climates of many islands, particularly those in the trade wind zone, where winds blow reliably from one direction almost all year.

The island of Hawaii is a particularly good example. Hawaii is composed of several volcanoes, some of which are over 4000 m high, and it is located in the northeast trade wind zone (**Fig. 7-33b**). Hilo, on the **windward** northeast coast, has rain almost every day. It is within a coastal belt of tropical rainforest, with many continuously flowing streams and waterfalls. Kailua-Kona, on the leeward southwest side of the island, is hot, sunny, and arid, with only occasional showers throughout most of the year. It is surrounded by almost bare fields of solidified **lava**, on which little can grow because of the low rainfall.

Other islands in the Hawaiian chain, and many other tropical Pacific islands, have windward wet climates and leeward dry climates. However, each island is somewhat different. On Oahu, for example, the mountains are relatively low and cut by passes through which clouds carry afternoon showers to Honolulu before they can dissipate. On other islands, the mountains are too low to produce the island effect. Kahoolawe, like most other low islands in the trade wind zone, is extremely arid and gets rainfall only from infrequent storms.

## 7.11: Chapter Summary

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### Atmosphere and Water Vapor.

Air moves vertically when a change in temperature or in its concentration of water vapor causes its density to change. In the lower atmosphere, vertical movements are generally limited to the troposphere (about 12 km altitude). The addition of water vapor to air decreases its density because lighter water vapor molecules displace heavier nitrogen and oxygen molecules. Atmospheric convection is caused by heating or evaporation at the sea or land surface. Water vapor continuously evaporates from the oceans into the atmosphere. As air rises, its pressure decreases, causing it to expand and cool. The saturation pressure of water vapor in air decreases with decreasing temperature. Hence, as air cools, it becomes supersaturated with water, which condenses to rain or ice crystals.

### Water and Heat Budgets.

Enough water is evaporated from land and oceans each year to cover the world about 1 m deep. About 93% of this water comes from the oceans, but nearly 30% of the resulting precipitation falls on land. The excess of precipitation over evaporation on land enters lakes, streams, and rivers and returns to the oceans as runoff.

Almost 25% of the solar radiation reaching the Earth is absorbed and converted to heat in the atmosphere. About 50% is absorbed by oceans and land, and the rest is reflected to space. Of the heat absorbed by the oceans, about half is lost by radiation, and half is transferred to the atmosphere as latent heat of vaporization. Solar energy per unit area received by the Earth is at a maximum at the equator and decreases toward the poles. However, the amount of heat radiated and reflected per unit area varies significantly with latitude and other factors. At the equator, more heat energy is received than is lost to space, whereas at the poles, more heat is lost than is received. Heat is transferred from the tropics to the polar regions by atmospheric and oceanic currents.

### Climatic Winds.

Ocean-to-atmosphere heat transfer, particularly through sensible and latent heat fluxes, plays a key role in fueling atmospheric convection, especially over tropical oceans. The atmospheric three-cell per hemisphere system consists of Hadley, Ferrel, and polar cells arranged between the equator and the pole in each hemisphere. This structure is modified, especially over the continents, as a result of land-ocean-atmosphere interactions driven by the far greater heat capacity of ocean waters compared to the land surface. Trade winds in the Hadley cell blow westward and toward the equator, and westerly winds in the Ferrel cell blow eastward and toward the pole. Between cells, winds are generally calm. Rainfall and clouds are heavy at upwelling regions, and winds are light and skies are clear at downwelling regions. The cells shift north and south seasonally as the Earth's angle to the sun changes. Because solar heat is released to the atmosphere relatively slowly by evaporation, the location of the atmospheric cells lags behind the location of greatest solar heating as the cells migrate seasonally.

### Climate and Ocean Surface Water Properties.

Ocean surface water temperatures generally decrease with latitude, but currents and upwelling distort the pattern. Surface water salinity is primarily determined by differences between evaporation and precipitation rates, except near continents, where freshwater runoff is high. Salinity is highest in the subtropics and polar regions, where rainfall is low. Salinity is low at mid latitudes, where rainfall is high and evaporation is less than at lower latitudes, and at the equator, where evaporation is reduced by persistent cloud cover and lack of winds. Extreme high salinity occurs in marginal seas where evaporation exceeds precipitation, and extreme low salinity occurs in marginal seas where precipitation exceeds evaporation.

### Interannual Climate Variations.

El Niño/Southern Oscillation is a complex, coupled ocean-atmosphere phenomenon that occurs across the equatorial region of the Pacific Ocean. Warm surface water is transported westward by the trade winds until it accumulates near Indonesia. When

the winds slow, the water flows back to the east along the equator, where there is no Coriolis Effect. During El Niño, upwelling is stopped near Peru, with often disastrous effects on marine life. Droughts occur in locations as far away as Central Europe, and severe storms occur in California and other places. When an El Niño ends, the system may overshoot to produce La Niña. However, the strength and severity of La Niña events vary, and in some regions La Niña can have impacts that are equally, or even more, pronounced than those of El Niño.

In addition to ENSO, interannual oscillations of ocean-atmosphere characteristics have been identified in the North Pacific, North Atlantic, Arctic, and Indian Oceans. Each of these oscillations affects regional climates, but the effects are generally smaller and less widespread than those due to ENSO. Studies of the Pacific Decadal Oscillation suggest that it, and probably other such oscillations, and their ecosystem effects are chaotic and may cause unpredictable irreversible ecosystem change.

## Global Climate Zones.

Climate zones in the oceans are generally arranged in latitudinal bands. Land climate zones are more complex and depend on proximity to the ocean and the locations of mountain ranges. The ocean moderates coastal climates because surface waters do not change much in temperature either during the day or during the year. The moderating influence of the oceans can extend far into the continents in regions where the winds are generally onshore and no mountain chains block the passage of the coastal air inland.

## Weather Systems.

In the free atmosphere, above the influence of surface friction, winds around high- and low-pressure systems often flow approximately geostrophically, or nearly parallel to isobars or height contours due to a balance between the pressure gradient force and the Coriolis force. Near the surface, however, friction disrupts this balance, causing winds to cross isobars toward lower pressure. Winds flow counterclockwise around a low and clockwise around a high in the Northern Hemisphere, and vice versa in the Southern Hemisphere. Tropical cyclones form over warm water in latitudes high enough that the Coriolis effect is significant. Tropical cyclone winds blow in toward the eye in a rising helical pattern. Winds accelerate as heat energy is added by evaporation from the ocean surface. but the hurricane loses strength when it moves over land or cool water, where its energy supply is removed. Extratropical cyclones form at the polar front as warm air flows over cold polar air and is deflected by the Coriolis effect.

## Local Weather Effects.

The Earth's surface heats by day from solar radiation and cools at night as the heat is lost. Primarily because of water's high latent heat, the ocean surface water temperature varies little in this daily cycle. Thus, land temperatures vary more. During the day, land next to an ocean heats, then loses some of its heat to the air. The warmed air rises and is replaced by cooler air from over the ocean, creating a sea breeze. At night, the land cools rapidly, which creates a land breeze that flows seaward to displace warmer, less dense air over the ocean. Coastal fogs form when warm, moist air from over the ocean passes over a cold coastal water mass as it enters the sea breeze system.

When moist air encounters mountains, it is forced to rise, expand, and cool. If it rises high enough, water vapor condenses or deposits onto nuclei and causes precipitation. After crossing the mountains, the air mass descends, is compressed, and warms. When it is no longer saturated with water vapor, precipitation ceases. Hence, the windward side of the mountains is wet, and the leeward side is arid.

## 7.11.1: Key Terms

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You should recognize and understand the meaning of all terms that are in boldface type in the text. All those terms are defined in the Glossary. The following are some less familiar key scientific terms that are used in this chapter and that are essential to know and be able to use in classroom discussions or exam answers.

- adiabatic expansion
- angle of incidence
- anthropogenic
- beach
- chaotic
- chlorofluorocarbons (CFCs)
- chlorophyll
- climate
- conduction
- convection
- convection cells
- convergence
- Coriolis effect
- crest
- cum sole
- currents
- cyclonic
- density
- diffusion
- diurnally
- divergence
- downwelling
- El Niño
- ENSO
- environment
- equinoxes
- erosion
- extratropical cyclones
- fetch
- food web
- friction
- fronts
- geostrophic
- heat capacity
- hurricanes
- ice exclusion
- intertropical convergence zone
- isotherms
- jet streams
- latent heat of vaporization
- latitude
- lava
- marginal seas
- marine mammals
- meanders
- nutrients
- overfished
- ozone layer
- phytoplankton

- pressure gradient
- pycnocline
- residence time
- salinity
- saturation pressure
- schools
- sediment
- sensible heat
- shores
- solstice
- spatial
- species
- storm surge
- subtropical high
- trade winds
- tropical cyclones
- troposphere
- turbidity
- water masses
- wave height
- weather
- westerlies
- windward

## 7.11.2: Chapter Questions

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### Study Questions

1. Why do the CFCs released into the lower atmosphere take a very long time to reach the Earth's ozone layer?
2. Why is the amount of water vapor in air so important?
3. What are the reasons for the net heat loss to space from polar regions of the Earth and the net heat gain from the sun in the tropics?
4. Why are there three atmospheric convection cells in the Northern and Southern Hemisphere? Discuss why the cells are not centered on the equator and why they are more complicated in the Northern Hemisphere.
5. Why are persistent cloud cover and rainfall present at atmospheric convergences over the oceans?
6. How do the locations of the world's deserts relate to the locations of the atmospheric convection cells?
7. Given that the summer solstice is in June, why is August usually the hottest month in most of the United States?
8. The Hawaiian Islands lie in the northeast trade wind zone. Why do the northeast sides of the islands have wet climates and the southwest sides dry climates?
9. Antarctica is a desert, despite its ice cap, which is several kilometers thick. Why is it a desert?
10. Why are the cloud patterns that we see in weather satellite images dominated by swirls?
11. Why don't hurricanes form at the equator? Why don't they form over land?

### Critical Thinking Questions

1. What causes air masses to move vertically in the atmosphere? Why is the vertical circulation of the lower atmosphere restricted to a layer of only about 12 km, although the entire atmosphere is more than 45 km deep?
2. Would the height of the restricted vertical circulation in the atmosphere be changed if the Earth's average surface atmospheric temperature increased? Why or why not?
3. On a nonrotating Earth, there would be one atmospheric convection cell in each hemisphere, with upwelling at the equator and downwelling at the poles. On the Earth today, we have three cells in each hemisphere because of the Coriolis deflection. Since the Coriolis deflection will be reduced as the Earth's rotation is progressively slowed down, at a lower speed of rotation would you expect a system with two convection cells in each hemisphere to be established eventually? Why or why not?
4. It is known that in Biblical times, parts of the areas that are now desert in Eurasia (Egypt, Israel, Syria, and so on) were fertile regions with plentiful rainfall. What does this tell us about the location of the atmospheric convection cells?
5. If the Earth's direction of rotation on its axis were somehow reversed, would the location of the world's deserts change? Why or why not? If so, which parts of each of the continents would become deserts, and why?
6. If the trade winds were strengthened by global warming, how might the frequency and strength of El Niños be affected? Explain the reasons for your answer.
7. If there were no mountain chains in the western part of the United States, how would the climate in Nevada and the plains states be different? Explain the reasons for your answer.
8. No hurricanes are fully formed on the North African side of the Atlantic Ocean, but many hurricanes are formed in the equivalent region off the Pacific coast of Mexico. Why?
9. Almost no hurricanes form in the South Atlantic Ocean. Why?

## 7.11.3: References

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