

ATMOSPHERIC PROCESSES AND PHENOMENA



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Atmospheric Processes and Phenomena

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This text was compiled on 05/19/2026

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About the authors

About the Contributors

This open access textbook was made possible through the collaboration of faculty, students, and staff at the University of Hawai‘i at Mānoa working together in the spirit of Aloha.

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None of this would have been possible without the help and guidance of Billy Meinke who served as project manager. His passion for Open Educational Resources is obvious, and guides his work and actions. Thank you Billy, your energy and devotion to OER is inspiring.

Acknowledgement

This Open Educational Resource textbook has been adapted from Roland Stull's OER textbook entitled "Practical Meteorology: An Algebra-based Survey of Atmospheric Science." It has been simplified and the content was reduced to meet the needs of the ATMO 200 course at the University of Hawai'i.

Chapters and section structures were adapted from the above existing OER textbook. Without this foundational text, a lot more work would have been required to complete this project. A huge thank you to Roland Stull.

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Special Thanks

Roland Stull — University of British Columbia ([link](#))

Bill Chismar — University of Hawai'i at Mānoa, Dean of the Outreach College

CHAPTER OVERVIEW

1: Atmospheric Basics

Alison Nugent

Learning Objectives

By the end of this chapter, you should be able to:

- Convert between temperature units of Fahrenheit, Celsius, and Kelvin
- Use mathematical formulas to define atmospheric temperature, pressure, and density
- Compute pressure and density changes with altitude
- Describe the vertical structure of Earth's atmosphere
- Define and apply the ideal gas law
- Describe hydrostatic balance
- Discuss the difference between weather and climate
- Note the location of terminology, coordinate systems, and units for future reference

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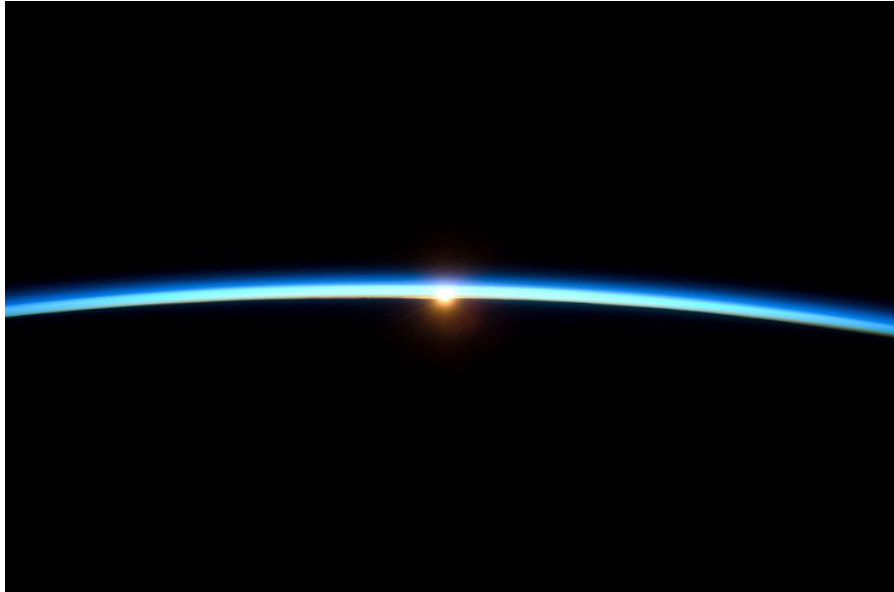
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1.1: Introduction

We often take Earth's atmosphere for granted. When we enjoy a sunset, go to the beach, or hike a trail, we don't necessarily think of the vast volume of air around us that allows us to do these things. The atmosphere brings us the oxygen that fills our lungs; it brings us the beautiful blues of the sky above and the verdant greenery below. It is responsible for the existence of our oceans, lakes, and rivers, and without it we would not have beaches to enjoy. The fluffy, white cumulus clouds you see on a summer day are a result of both large- and small-scale motions in the atmosphere. The same is true of the hazy stratus, wispy cirrus, or towering cumulonimbus. The atmosphere is like a massive blanket that surrounds, sustains, protects, and warms us. Without it, no life would exist on Earth. In fact, the atmosphere is the very reason almost anything exists on Earth at all; our planet would be a dull, waterless, lifeless rock without it. Night-time without an atmosphere would be unimaginably cold, and daytime temperatures would soar above water's boiling point. Nothing would separate Earth and the blistering sunlight but the empty vacuum of space.



View from outer space of the sun rising over Earth, illuminating the atmosphere in a ring of blue. [NASA](#) (Public Domain).

Although the atmosphere was previously described as “vast”, in actuality it is relatively thin. If the Earth were shrunk down to the size of a basketball, the atmosphere would be roughly the thickness of a plastic sheet stretched across the ball. In the image above, the atmosphere can be seen as the thin veil of bluish white mist above our planet's surface. Although we can travel thousands of kilometers horizontally along Earth's surface, traveling vertically would be difficult — the air is too thin to breathe only a few kilometers above Earth's surface.



View of the moon from space over Earth's blue atmosphere ([Public Domain](#)).

The goal of this chapter is to orient those who are new to atmospheric science with the basics as described previously in the Chapter 1 Learning Objectives.

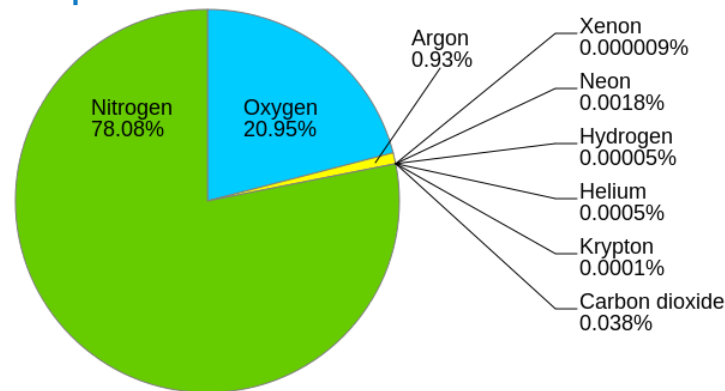
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1.2: Overview of Earth's Atmosphere

Everything that happens on Earth is caused in some way by radiation from the sun, which is an average-sized star located near the edge of the Milky Way galaxy. It provides radiant energy (also called radiation or solar insolation) by converting hydrogen into helium near its core, which provides most of the Earth's warmth. Here is a [classic song](#) about the Sun heating Earth.

The Earth only receives a tiny portion of the sun's total energy output. It is this radiation that drives the atmosphere's wind and weather patterns on the Earth's surface. It is because of this solar input that the Earth can maintain an overall global average surface temperature of approximately 15°C (59°F).

Composition of the Atmosphere



Pie chart showing the gases that make up Earth's atmosphere ([CC BY-SA 3.0](#)).

Earth's atmosphere is primarily composed of nitrogen (N) and oxygen (O), with smaller quantities of other gases, as shown in the figure above. Nitrogen makes up around 78%, oxygen makes up around 21%, and Argon almost 1% of the total dry air volume in the atmosphere. Other gases (such as water vapor, the gaseous form of H₂O) take up varying amounts depending on the location and atmospheric conditions. There is a balance of input and output of atmospheric gases at the Earth's surface both from life and surface processes. For example, when we breathe, the human body takes in oxygen (O₂) and releases carbon dioxide (CO₂). When plants photosynthesize, they take in carbon dioxide and release oxygen. When water (H₂O) evaporates from the ocean, additional water vapor (H₂O) is added to the atmosphere. When the Kilauea volcano degasses, sulfur dioxide (SO₂) is added to the atmosphere.

Water Vapor

Water vapor concentrations vary greatly from location to location, and from time to time. In warm, moist, tropical regions, water vapor can make up nearly 4% of the atmospheric composition, while in polar regions it can be just a tiny fraction of a percent. Water vapor cannot be seen in its gaseous form, but you may see it as it *condenses* into water droplets on the side of a cool glass of water on a hot day or into water droplets that make up clouds. The process of water vapor changing phase to liquid is called *condensation*. When liquid water becomes a gaseous vapor, it is known as *evaporation*. When water becomes ice or ice becomes water, these processes are known as *freezing* and *melting*, respectively. When ice transitions directly to water vapor, the process is known as *sublimation*. The transition from water vapor to ice is known as *vapor deposition*. These terms will all be used frequently in subsequent chapters.

Water vapor is one of the most important gases in our atmosphere, because when it changes phase from vapor to liquid or ice, it releases enormous amounts of heat, called **latent heat**. Latent heat is a major source of energy in the atmosphere, particularly for tropical storms and other types of convection. In addition, water vapor is an important greenhouse gas in the atmosphere, meaning it absorbs and re-releases a portion of the Earth's outgoing radiation, which allows our planet to remain warm.

If you don't understand convection, latent heat, and greenhouse gases yet, don't worry, all will be discussed in later chapters. The point is that despite its small fractional percentage, water vapor is arguably the most important gas in the atmosphere.

Hawaiian Focus Box

The air composition of Hawai'i occasionally faces a unique threat due to the emissions from Kīlauea, a volcano located along the southern shore of Hawai'i's Big Island. Kīlauea has been continuously erupting since 1983 and is the most active of the volcanoes that make up the Big Island. In the first few years after 1983, Kīlauea emitted up to 30,000 tons of sulfur dioxide (SO_2) a day, but that number has stabilized recently to around 5,000 tons per day. At present day with the 2018 Kīlauea eruptions currently occurring, SO_2 emissions are again elevated. These emissions are known as volcanic smog, or more commonly, "vog". The sulfate particles in vog are very small, so they can effectively infiltrate human airways. In environments with high relative humidity, such as inside the human body, these particles will hydrolyze and expand, which irritates lungs and obstructs airways. Vog has been linked to respiratory disease, sinusitis, and even lung cancer. Therefore, Hawai'i's residents need to take precautions.



Water vapor and other gases escaping from the Kīlauea volcano at the Halema'uma'u vent ([CC BY-SA 4.0](#)).

Vertical Structure of the Atmosphere

Although the atmosphere extends vertically for hundreds of kilometers, almost 99% of it is within approximately 30 km of the Earth's surface. Air molecules are pulled toward Earth by the gravitational force, which pulls downward on the atmosphere. This causes the air molecules closer to Earth's surface to be more tightly compressed, meaning that there are more air molecules together in a given volume (a higher density). The greater the number of air molecules that exist above a certain altitude, the greater the effect of this compression, because the molecules are all being pulled downward together. Just as gravity has an effect on the weight of different objects (weight is the force that acts on an object due to gravity), it also gives weight to air.

$$\text{Weight} = \text{mass} \times \text{gravitational acceleration}$$

$$\text{Weight} = m \times g$$

Mass is the measure of how much matter exists in a given object or space. Mass is occasionally confused with weight. While weight increases with more mass, mass would have no weight at all without the pull of gravity. Similarly, an object's weight depends on the gravitational force: an object weighs much less on the moon than it does on Earth. Mass is typically given in grams (g) or kilograms (kg).

Note

You may want to review the International System (IS) of Units. The base units we will use include meters (m), kilograms (kg), seconds (s), and Kelvin (K). We'll also use the prefixes "kilo" (1000), "hecto" (100), and "micro" (10^{-6}) along with some derived units like hertz, Newtons, pascals, joules, and watts.

Air Density

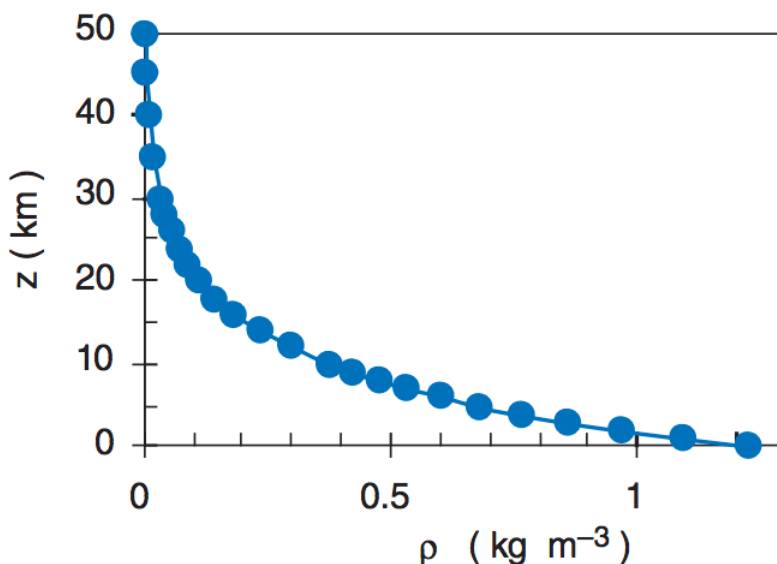
Air density is determined by the amount of mass (m) that exists in a given space, or volume (V). If you fit a lot of mass into a tiny volume, you will have higher density. If you have only a small amount of mass spread out over a large volume, the density will be lower.

$$\text{Density} = \frac{\text{mass}}{\text{volume}}$$

$$\rho = \frac{M}{V}$$

Putting these ideas together, air density is highest near the surface of Earth and decreases with height, because there are more molecules held tightly together at the surface by gravity than there are above. The standard unit for density is $\text{kg}\cdot\text{m}^{-3}$.

Pro Tip: Density is often denoted by the Greek lowercase letter ρ (rho). Be careful not to confuse this with p or P , which are both used to denote pressure. When writing on paper, it may be useful to write ρ with a curly tail like a backwards q and to write p with a straight tail in order to differentiate the two.



Density (ρ , $\text{kg}\cdot\text{m}^{-3}$) vs. height (z , km) in the atmosphere ([CC BY-NC-SA 4.0](https://creativecommons.org/licenses/by-nc-sa/4.0/)).

The image above shows the distribution of density (ρ) with altitude. Atmospheric surface density is typically around $1.2 \text{ kg}\cdot\text{m}^{-3}$. We call this value ρ_0 because it is the initial value at the surface. The exponential equation below approximates the distribution of density with height

$$\rho(z) = \rho_0 \times \exp(-z/H)$$

$$H = \frac{R_d \times T}{g}$$

where ρ (kg m^{-3}) is density, ρ_0 (kg m^{-3}) is the density at the Earth's surface, z (m) is altitude, T (K) is temperature (assumed to be constant through the atmosphere), $R_d = 287.053 \text{ J}\cdot\text{K}^{-1}\cdot\text{kg}^{-1}$ (gas constant for dry air), and g is the acceleration due to gravity ($\text{m}\cdot\text{s}^{-2}$).

 Note

Remember that the equations you see in this text are good approximations. They all involve assumptions of some kind. In the case of the exponential distribution of density with altitude, the primary issue is the assumption that temperature is constant throughout the atmosphere. This will become clear later.

“Scale height” is denoted by H and represents an e-folding distance for the drop off of density in the atmosphere. Typically the value of H is around 8000 m. This means that density at 8 km is approximately $1/e$ ($e=2.71828$) of the value at the surface. A back of the envelope calculation gives the density of the atmosphere at 8 km as $1.2 \text{ kg}\cdot\text{m}^{-3}$ divided by 3 and is approximately equal to $0.4 \text{ kg}\cdot\text{m}^{-3}$. Checking the image above, it is clear that this provides a relatively good estimate for density.

Air Pressure

You will often hear meteorologists on TV discuss the air pressure, or you might see H’s and L’s on a map displayed on the Weather Channel, denoting high and low pressure areas. Because air molecules are in constant motion, they collide with one another and other objects up to several billions of times a second. Each time an air molecule collides with an object, it exerts a tiny amount of force. Air pressure refers to the total force that air exerts against a given area of an object.

$$\text{Pressure} = \frac{\text{Force}}{\text{Area}}$$
$$P = \frac{F}{A}$$

A Newton (N) is the unit for force and m^2 is the unit for area, in the International System of Units (SI). Therefore, the standard unit for pressure is in Newtons per square meter — or Pascals (Pa), which are defined as $1 \text{ N}\cdot\text{m}^{-2}$.

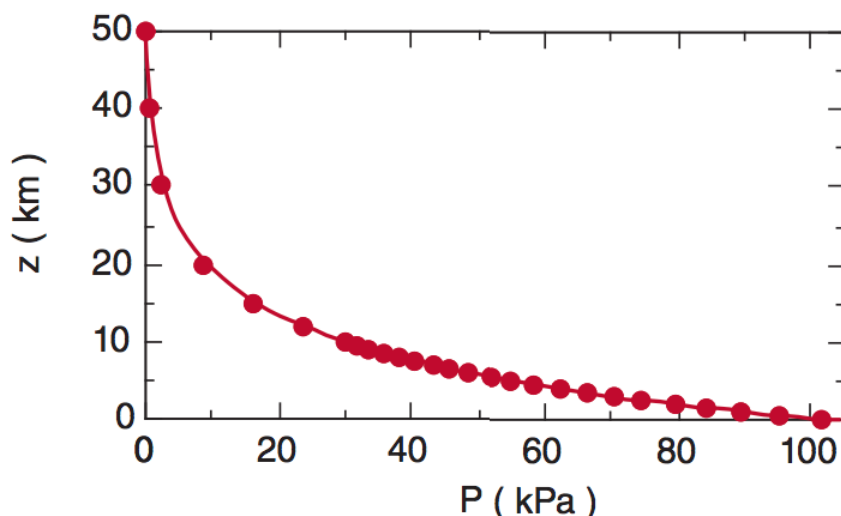
Pro Tip: You will generally find air pressure expressed in units of millibars or (mb) or hectopascals (hPa) on weather maps. These two units are equivalent to one another. In the *Practical Meteorology: An Algebra-based Survey of Atmospheric Science* textbook by Roland Stull off of which this OER is based, kPa are often used. In the aviation field, Inches of mercury (inHg) are also commonly used. At sea level, the global average for atmospheric pressure is:

$101.325 \text{ kPa} = 1013.25 \text{ mb} = 1013.25 \text{ hPa} = 29.92 \text{ in. Hg.} = 1 \text{ atm (atmosphere)} = 101325 \text{ Pa.}$

In future calculations, you will usually need to express pressure in Pascals (Pa) for your units to cancel out. Always be mindful of the units you are given and be sure you are able to convert from one type of unit to another.

You can think of the surface air pressure as the total weight of a column of air molecules extending from the surface to the top of the atmosphere.

Because there are more air molecules at the surface of the Earth and less above, air pressure is maximized at the surface and decreases with height nearly exponentially, analogous to air density.



Height (z , km) vs. pressure (P , kPa) in the atmosphere ([CC BY-NC-SA 4.0](#)).

The image above shows the distribution of pressure (P) with altitude. Atmospheric surface pressure is typically around 1013 hPa. We call this value P_0 because it is the initial value at the surface. The exponential equation below approximates the distribution of pressure with height

$$P(z) = P_0 \times \exp\left(\frac{-z}{H}\right)$$

$$H = \frac{R_d \times T}{g}$$

where P (Pa) is pressure, P_0 (Pa) is the pressure at the Earth's surface, z (m) is altitude, T (K) is temperature (assumed to be constant through the atmosphere), $R_d = 287.053 \text{ J}\cdot\text{K}^{-1}\cdot\text{kg}^{-1}$ (gas constant for dry air), and g is the acceleration due to gravity ($\text{m}\cdot\text{s}^{-2}$).

You can get a sense of atmospheric pressure if you've ever dived more than a few feet underwater at the pool or beach. As you dive deeper, the weight of the water above you increases, and you feel increased discomfort or pressure on your head. This is why divers require special equipment to reach greater depths in the ocean. If you think of yourself at the bottom of an ocean of air, you can imagine why air pressure is highest down here at the surface. If you've ever flown on an airplane, you can also feel the changes in atmospheric pressure in your ears as you ascend and descend. Still, while pressure decreases at high elevations in an airplane, cabins are pressurized to maintain similar pressure to Earth's surface so you are not experiencing the full pressure drop with altitude.

Air Temperature

It is commonly thought that air temperature is a measure of how cold or hot an object is. We instinctively know that the higher an object's temperature, the hotter it will be, and the lower its temperature, the colder it will be. But what does temperature really measure? The answer lies in the motion of the molecules.

Air molecules are in constant motion, colliding with objects and one another. As we increase the temperature of a volume of air, the air molecules speed up, and collisions become more frequent. As we decrease the air temperature, the molecules slow down, and collisions occur less frequently. What is going on here? Simply put, air temperature is a relative measure of the average kinetic energy (kinetic meaning it relates to motion) of the molecules of a system.

Because there are more air molecules close to the Earth's surface, density and air pressure are maximized at the surface and decrease with height. Based on this, do you think air temperature will decrease or increase as you move further away from the surface?

Within the lowest 10-12 km of the atmosphere, temperature tends to decrease with height, primarily because the Earth's surface is warmed by sunlight, which then warms the layer of air directly above it, which warms the air above that, and so on. While this is true in the lowest surface layer of the atmosphere, this is not true throughout the rest of the atmosphere. The vertical temperature

profile is more complicated than the vertical profile of pressure or density. Temperature is different for each of the different layers of the atmosphere, which will be discussed later.

Temperature Scales

Surface air temperature is commonly given in degrees **Fahrenheit** (°F) in the United States. If you were born and raised in the US, this is the temperature scale that you probably use in your daily life, and what you'll see given when you look at the daily weather. You'll know that 32°F is the temperature at which water freezes, and 212°F is the temperature at which water boils.

Degrees **Celsius** (°C) is also commonly used, especially internationally. Celsius is a convenient scale to use because water freezes at 0°C and boils at 100°C. A difference in 1°C is larger than a difference of 1°F by about 1.8 times. To convert between the two, use the following equations.

$$^{\circ}C = \frac{5}{9} \times (^{\circ}F - 32)$$

$$^{\circ}F = \left(\frac{9}{5} \times ^{\circ}C \right) + 32$$

The **Kelvin** scale (K) is almost always used as a temperature unit in scientific equations and is convenient in that it contains no negative numbers. The Kelvin scale begins at 0 K, or absolute zero, where atoms and molecules would theoretically be thermally motionless. The Kelvin scale is also sometimes known as the absolute temperature scale. The lowest possible temperature is 0 K but it does not occur naturally. The coolest naturally existing place known in the universe is the Boomerang Nebula, located in the Centaurus constellation about 5,000 light-years away from Earth. The temperature is measured at 1 K, only 1°C above absolute zero. A difference in 1 K is the same as a difference of 1°C, so a conversion is linear and simple.

$$K = ^{\circ}C + 273.15$$

Based on this, absolute zero is -273.15 °C. Keep in mind degrees Celsius (°C) and degrees Fahrenheit (°F) always have the degree ° symbol in front, but Kelvin (K) never uses this symbol. Typical values of temperature on Earth's surface on the Kelvin scale are values around 260-310 K.

Within atmospheric sciences, the Kelvin and Celsius scale are used. This chapter is one of the only times you'll see a discussion of Fahrenheit within the course.

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1.3: Equation of State — Ideal Gas Law

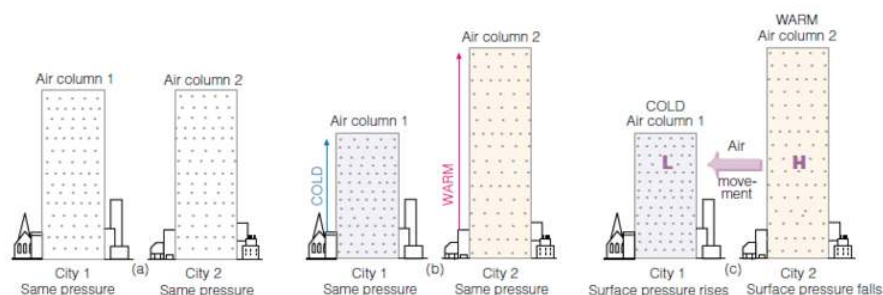
Air pressure is caused by the collisions of rapidly, randomly moving air molecules, so you might expect pressure to increase when there are more molecules in one place (higher density, ρ) and when they are moving faster (higher temperature, T). The relationship between pressure, density, and temperature is called the **Equation of State**.

The gases in the atmosphere have a simple equation of state called the **Ideal Gas Law**.

For dry air (no water vapor present), the ideal gas law is

$$P = \rho \times R_d \times T$$

The ideal gas law assumes that atmospheric gases act in an ideal manner, meaning that there are few intermolecular forces, and that the size of the molecules is small compared to the space between them. In almost all cases, gases in the atmosphere can be assumed to be ideal gases. Again, this is another example of a simplifying assumption we use to be able to approximate relationships in the atmosphere.



Air pressure differences associated with air temperature in a column ([CC BY-SA 4.0](#)). Description below.

The above figure illustrates the effect that air temperature has on its density and pressure. The two air columns above Cities 1 and 2 in (a) have the same temperature, contain the same amount of mass, and have the same volume, so their density is the same. The pressure exerted on the surface is the same in both cities because the pressure at the surface is related to the number of air molecules above.

In (b) the air temperature above City 1 is lower than the air temperature above City 2. Because of this the air molecules in the column above City 1 are moving more slowly, and take up a smaller volume. Likewise, in City 2 the air molecules are moving more quickly and the volume in the column is larger. It only takes a smaller cold air volume to apply the same pressure on the surface as a larger warm air volume. The surface pressures are the same but the temperatures are not.

The surface pressure may be the same in (b) but the pressure aloft is not, which results in air movement that can be seen in (c). At the same altitude above Cities 1 and 2 in (c), there is more air above the same level in City 2 than in 1. Locally this creates a higher pressure aloft over the warmer City 2. Because of the difference in pressure aloft, airflow is created that moves from the higher pressure toward the lower pressure. This airflow caused by a localized difference in air pressure is what we call wind.

Note

The force that arises due to a difference in air pressure is called the **pressure gradient force (PGF)**. The word “gradient” here refers to a change over a distance. A pressure gradient force is the force that drives the wind. This concept will be further explored in Chapter 10.

The movement of air from City 2 to City 1 creates a falling surface pressure in City 2 and a rising surface pressure in City 1.

1.3: Equation of State — Ideal Gas Law is shared under a [CC BY-SA 4.0](#) license and was authored, remixed, and/or curated by LibreTexts.

1.4: Hydrostatic Balance

We've learned that atmospheric pressure decreases with height. We've also learned that air moves from areas of high pressure to areas of low pressure due to the pressure gradient force. Knowing these two things, you might think that the air in the atmosphere would escape into space, because there is high pressure at the surface and low pressure aloft. Because it does not, there must be a downward force that balances the upward vertical pressure gradient force. This downward force is a force that should be very familiar to you: gravity.

In the atmosphere, when the vertical pressure gradient force is balanced by gravity, this is known as **hydrostatic balance**. The word hydro means water or fluid, and static means stationary, so the name can be interpreted as a stationary fluid balance. This balance holds true for most situations in the atmosphere. The hydrostatic equation is given by

$$\frac{\Delta P}{\Delta z} = -\rho * |g|$$

where $g = -9.8 \text{ m}\cdot\text{s}^{-2}$ is the acceleration due to gravity.

The negative sign here is due to the fact that pressure is decreasing as height increases so the left-hand side will be negative.

If you plan to use the above equation to calculate changes in altitude (Δz) with changes in pressure (ΔP) or vice versa, note that the above equation applies best over small changes. If the change in pressure or altitude is large, the exponential equation for $P(z)$ defined above is best.

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1.5: Hypsometric Equation

You have learned that air pressure decreases with height, but sometimes we want to know the distance or thickness between two different pressure levels. The atmospheric **thickness** varies depending on the average temperature in the layer. Warmer air is spaced out more, so a warmer layer of air will be thicker, while a cooler layer of air will be thinner. By knowing the average temperature of the layer, and the top and bottom pressure levels, you can calculate the thickness of the atmospheric layer. The hypsometric equation allows you to calculate how pressure varies with height in an atmosphere regardless of the temperature profile. It is the result of combining the ideal gas law with the hydrostatic equation. The **hypsometric equation** allows you to calculate the thickness ($z_2 - z_1$) between two pressure levels, P_2 and P_1 . The z_2 and z_1 values are the heights at pressure levels P_2 and P_1 , respectively.

$$z_2 - z_1 \approx \frac{R_d}{g} * \bar{T} \times \ln \left(\frac{P_1}{P_2} \right)$$

$$P_2 = P_1 \times \exp \left(\frac{z_1 - z_2}{\frac{R_d}{g} \times \bar{T}} \right)$$

In the above equation, the average temperature is shown. If the atmosphere is very moist, you may wish to use the average virtual temperature between the two heights z_2 and z_1 instead which includes the effects of water vapor.

T_v is known as the **virtual temperature**, defined as:

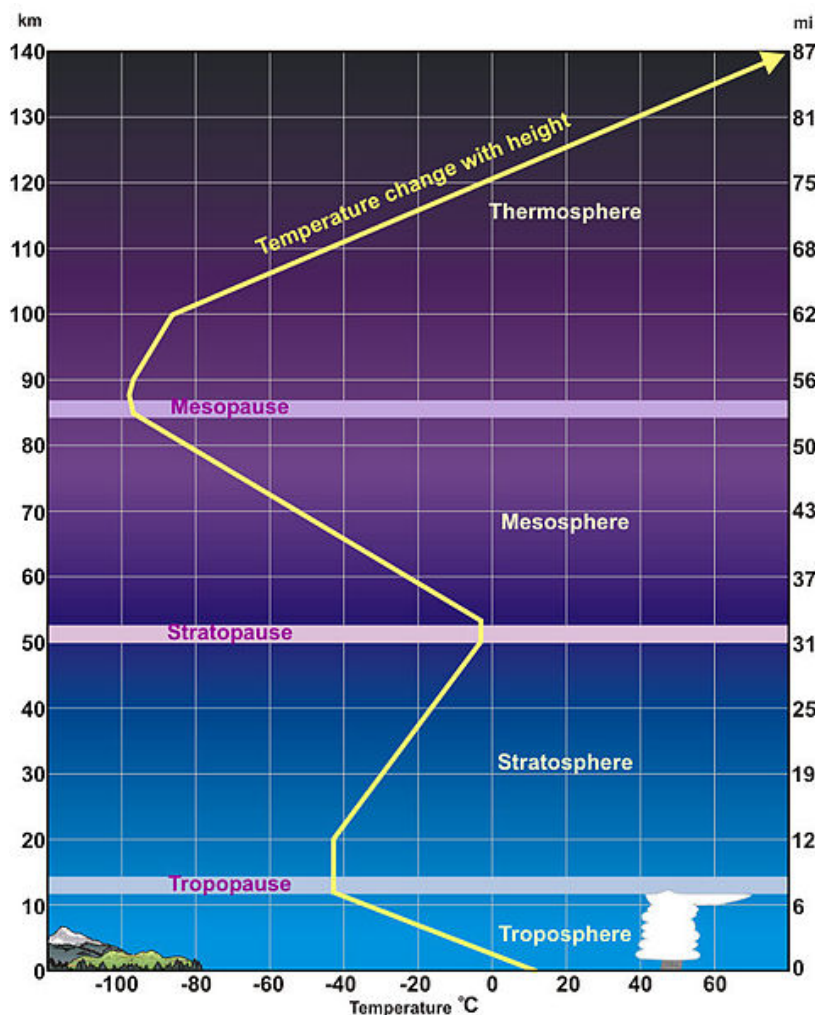
$$T_v = T \times [1 + (0.61 \times r)]$$

where the mixing ratio (r) is the mass of water vapor per mass of dry air and uses units of kilograms of water vapor per kilograms of dry air ($\text{kg} \cdot \text{kg}^{-1}$). You will learn that many times virtual temperature can be used inside of equations in place of temperature if the effect of water vapor needs to be included.

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1.6: Layers of the Atmosphere

In the previous sections, we talked about how both air pressure and density decrease with height, because most of the atmosphere is held close to the Earth's surface. This change in pressure and density occurs quickly at first, but slows down at higher altitudes. Unfortunately the change in temperature with altitude is not nearly as straightforward. When we look at a vertical profile of the atmosphere, we see that it can be divided into different layers in a number of ways. We can define layers by how the air temperature varies with height, by the gas composition, and even by the electrical properties of each layer.



The vertical temperature structure of Earth's atmosphere ([Public Domain](#)) in kilometers (left axis) and miles (right axis). The layers are labeled as "spheres" and the boundaries between the layers are labeled as "pauses". For example the troposphere is the lowest layer and the tropopause is the boundary between the troposphere and the stratosphere. The yellow line shows temperature in °C, and the cloud indicates where most of the weather occurs.

As can be seen in the above figure, the air temperature decreases with height up to the tropopause, 11 km in this example. This occurs because radiation from the sun warms the Earth's surface, and the surface warms the air just above it. This will be discussed in further detail in later chapters.

Troposphere

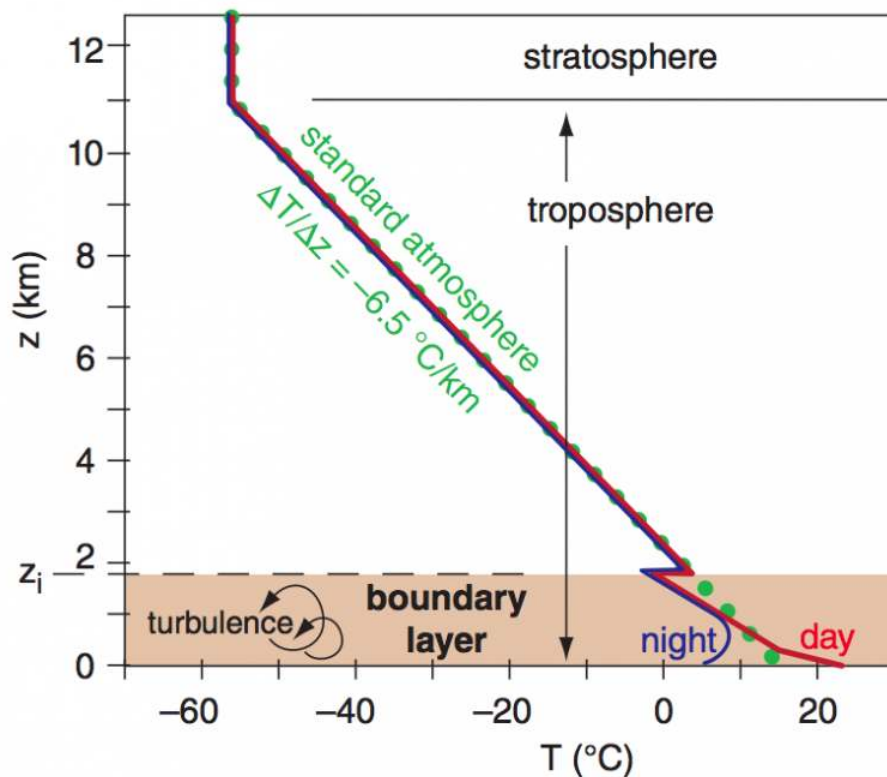
The layer of the atmosphere that we are most familiar with is the troposphere, which extends from the surface to around 11 km. All of the every day weather that we experience on Earth happens within the troposphere, which is characterized by frequent rising and sinking vertical air motions. The troposphere gets its name from the root word "tropo" in Greek, which means turning.

At the top of the troposphere, the air temperature stops decreasing with height in a region known as the tropopause, which separates the troposphere and the stratosphere above. The height of the tropopause varies depending on the season and location. In warmer

areas near the equator, the tropopause tends to be higher (around 17 km), while in colder polar regions the tropopause is lower (around 9 km) because warm layers of air are thicker than layers of cold air. For the same reason, the tropopause is found at higher elevations in the summer, and at lower elevations in the winter. Aircraft fly at the tropopause height.

Atmospheric Boundary Layer

The **atmospheric boundary layer** (ABL) is located within the lowest 0.3 to 3 km of the troposphere and is affected in several ways due to its close proximity with the Earth's surface. Airflow closest to the surface slows down due to the effect of friction, which can be amplified due to the type of terrain or amount of vegetation. Because of this, the boundary layer experiences the greatest amount of turbulence in the atmosphere. In addition, the boundary layer is warmed by the Earth's surface during the daytime, so it is the layer that is most affected by the diurnal (day-night) heating cycle.



The change in temperature with altitude in the troposphere which is the lowest layer of Earth's atmosphere. The turbulent boundary layer near the surface is also seen (shaded tan), along with the standard atmospheric lapse rate ($-6.5\text{ K}\cdot\text{km}^{-1}$), and near-surface day (red) / night (blue) temperature differences ([CC BY-NC-SA 4.0](https://creativecommons.org/licenses/by-nc-sa/4.0/)).

Stratosphere

In the *stratosphere*, the air temperature increases with height, causing a temperature inversion. This inversion layer tends to keep rising and sinking tropospheric air from mixing with stratospheric air. It also prevents rising and sinking from happening in the stratosphere itself. Because of this, it is called a *stratified* layer. The root word “strato” means layered, or spreading out, which is a good way to describe the many layers of the stratosphere. The temperature of the stratosphere increases with height because of the presence of a gas called ozone (O_3), which heats the air through the absorption of ultraviolet (UV) radiation from the sun.

At around 50 km, where the stratosphere is warmest (due to most of the UV radiation absorption occurring in the uppermost parts of the layer). The boundary known as the *stratopause* separates the stratosphere below from the mesosphere above.

Mesosphere

In the *mesosphere*, the air temperature once again decreases with height and the air is extremely thin with a low density. The average atmospheric pressure in this layer is about 1 mb, which means that about 99.9% of all air molecules are located below this level and only about one thousandth of the atmospheric mass is located above. You would not survive very long in the mesosphere,

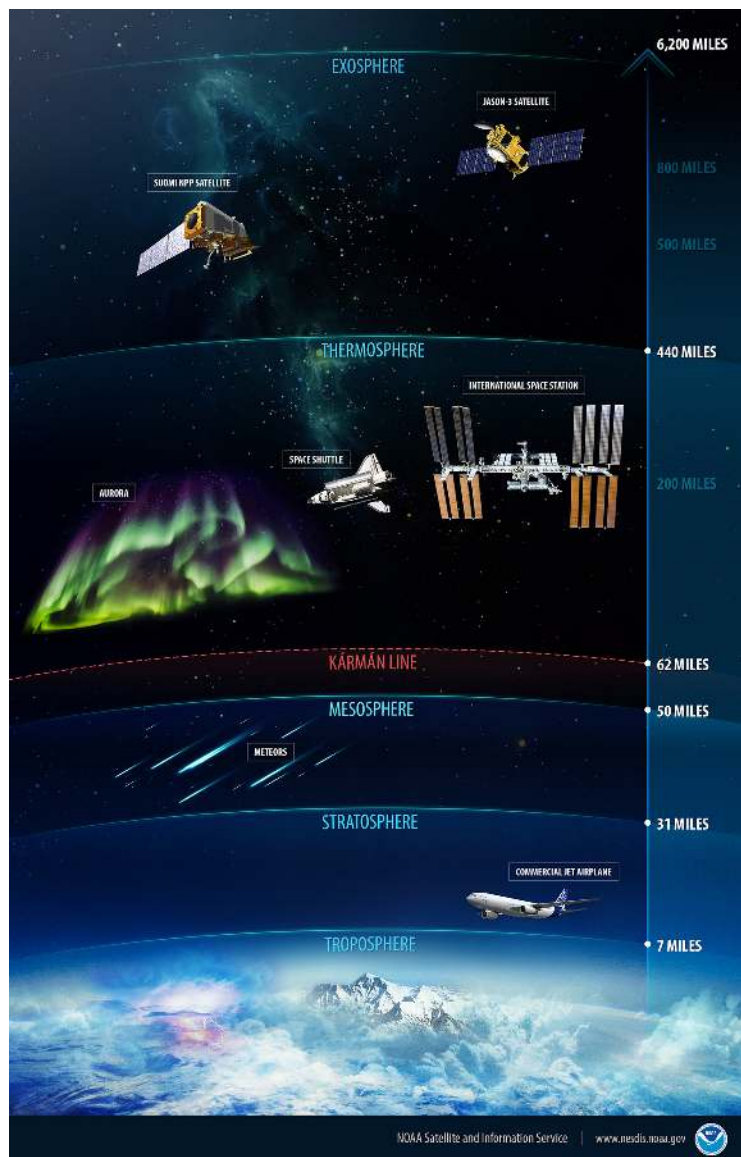
however, due to the extreme thinness of the air, freezing temperatures, and direct exposure to ultraviolet radiation. In addition, your blood would begin to boil at normal body temperatures if it were directly exposed to the low air pressure. The reason that temperatures decrease with height in the mesosphere is due partially to the fact that there is not a lot of ozone to absorb UV radiation. Because of this, the air molecules in the mesosphere lose more energy than they absorb.

At about 85 km near the *mesopause*, the atmosphere is at its coldest at about -90°C . The mesopause separates the mesosphere from the thermosphere above.

Thermosphere

In the *thermosphere*, the air density is extremely low so even a small amount of radiation absorption can lead to a large increase in temperature. Molecules can travel for entire kilometers without colliding into another molecule. This is the layer where auroras occur as a result of interactions between charged solar particles and air molecules.

At the top of the thermosphere, at above 500 km above the Earth's surface, molecules can actually escape the gravitational pull of the Earth. This region is known as the *exosphere*, and represents the upper limit of the atmosphere.



This figure illustrates the relative heights of different phenomena in the various vertical layers of the atmosphere. The Kármán Line at 62 miles (100 km) designates the boundary between Earth's atmosphere and outer space. Image by [NOAA Satellites \(Public Domain\)](#).

The previous figure shows the levels in Earth's atmosphere and the phenomena and activities that occur there. For example, commercial jet airplanes fly at the base of the stratosphere, while the International Space Station is positioned in the thermosphere.

Chapter 1 contained a vast array of topics, from defining temperature and pressure, to describing atmospheric vertical structure and components. As a reminder, these were our learning goals:

1. Convert between temperature units of Fahrenheit, Celsius, and Kelvin
2. Use mathematical formulas to define atmospheric temperature, pressure, and density
3. Compute pressure and density changes with altitude
4. Describe the vertical structure of Earth's atmosphere
5. Define and apply the ideal gas law
6. Describe hydrostatic balance
7. Discuss the difference between weather and climate
8. Note the location of terminology, coordinate systems, and units for future reference

The next section will discuss weather vs. climate and various terms and coordinate systems that are useful to review.

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1.7: Weather and Climate- What's the Difference?

Occasionally the terms “weather” and “climate” get confused with one another. Climate change has been a hot topic in recent decades. Knowing this, you might conclude that when the climate changes — it must be a pretty big deal, however, the weather is in a constant state of change. If it weren't, we would have little need to check the Internet or local weather station to see what the sky is going to look like tomorrow. Nobody bats an eye when the weather is different from one day to the next. What do we mean exactly by the terms weather and climate?

Weather refers to the present condition of the atmosphere at any given time and place. This includes the following elements.

Air Temperature — how hot or cold the air is.

Air Pressure — the force the air exerts on the surface.

Humidity — the amount of water vapor present in the air.

Clouds — masses of water droplets and/or ice crystals that obscure parts of the sky.

Precipitation — water (solid or liquid) that falls from clouds and reaches the ground

Visibility — the maximum horizontal distance that can be seen. This can be affected by the presence of fog or precipitation.

Wind — the horizontal flow of air, caused by local differences in air pressure.

Climate represents the average range of weather events in a region over a long period of time including weather extremes such as heat waves or cold spells, as well as the frequency of these events.

One way to think of it is weather influences the kind of clothing we might wear that day, e.g., Will I need a raincoat? Is it too warm to wear long sleeves? Is it too cool to wear shorts? Climate, on the other hand, influences what clothing we buy. For example, in Hawai'i, it isn't likely that you would need to buy a heavy winter jacket.

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1.8: Chapter 1 Reference Guide- Coordinate Systems, Units, Terminology

This reference guide covers some topics that are outlined in Roland Stull's *Practical Meteorology: An Algebra-based Survey of Atmospheric Science* Chapter 1, which may be useful to you going forward. These topics include the following.

Meteorological Conventions

This section describes standard meteorological conventions that are necessary to keep in mind when solving problems in this course. It includes different coordinate systems, including Cartesian coordinates, polar coordinates, and spherical coordinates. You are likely already well acquainted with Cartesian coordinates, but polar and spherical coordinates may be new to you. The standard way of describing and plotting wind direction is also given here.

Cartesian coordinates: x, y , and z and velocity: u, v , and w .

Polar coordinates: direction and magnitude.

Wind in Cartesian coordinates: (U, V)

Wind in Polar coordinates: (α, M)

Algebraically, wind is given a magnitude (wind speed) and a direction. Wind direction is split up into different components. When only horizontal wind motion is taken into account, we use U and V , and vertical air motions are denoted by W .

$U \rightarrow$ Wind in the x -direction

$V \rightarrow$ Wind in the y -direction

$W \rightarrow$ Wind in the z -direction (vertical air motion)

Because of this, wind can be plotted using polar coordinates, wind direction (dir) and wind magnitude (spd).

Wind directions are given by angle, with 0° to the north, and degrees increasing clockwise. **Winds are described using the direction from which they come.** A westerly wind is a wind from the west. Hawai'i is often impacted by the easterly or northeasterly trade winds, which come from the east and northeast.

$$spd = (U^2 + V^2)^{1/2}$$

$$dir = 90 - (360/C) * \arctan\left(\frac{V}{U}\right) + 0$$

Sometimes, cylindrical coordinates (M, α, W) are used which are similar to polar coordinates in that magnitude and direction of wind velocity are used, but also include the vertical motion component W .

The Stull text uses the words “ordinate” and “abscissa”. Ordinate is simply the vertical axis (typically the y -axis) and abscissa is the horizontal axis (typically the x -axis). Independent variables are usually plotted on the x -axis, with dependent variables plotted on the y -axis.

Earth Frameworks Reviewed

The Earth is not a sphere, but it is pretty close. The distance between the center of the Earth and the north and south poles differs by about 20 km, so the Earth is referred to as an oblate spheroid.

Cartography

Meridians, which are north-south lines on a globe, are given by degrees longitude. Think of the distance between the north and south poles as long.

The **prime meridian** lies at 0° longitude, and passes through Greenwich, Great Britain.

East of here (defined as the Eastern Hemisphere, $0 - 180^\circ \text{E}$), longitude is positive.

West of the prime meridian (Western Hemisphere, $0 - 180^\circ \text{W}$), longitude is negative.

The Earth rotates counterclockwise about its axis.

Parallels, which are east-west lines on a globe, are given by degrees latitude. A good way to remember this is “Lat” rhymes with “flat” — just like the east-west horizontal (flat) lines of latitude.

The Equator is 0° latitude, with latitudes north of the Equator as positive (Northern Hemisphere: $0 - 90^\circ\text{N}$), and latitudes south of the Equator are negative (Southern Hemisphere: $0 - 90^\circ\text{S}$).

A helpful approximation between degrees of latitude and distance is that each degree of latitude is approximately 111 km, or about 60 nautical miles.

Chapter 1: Questions to Consider

1. Drag and drop the correct labels to the appropriate layers of the atmosphere.
2. Mauna Kea on Hawai'i Island is 13,803 feet tall at the summit. If the pressure at sea level is 1015 hPa and the scale height is 8 km, what is the pressure at the summit? (Hint: don't forget to check the units!)
3. The pressure at the ground floor of a high-rise building is 1012 hPa. On the roof the pressure is 1007 hPa. If you assume that the air density is $1.2 \text{ kg}\cdot\text{m}^{-3}$ and acceleration due to gravity is $-9.8 \text{ m}\cdot\text{s}^{-2}$, approximately how tall is the building?
4. In this chapter you learned about two different equations that describe the change in air pressure (P) with altitude (z):

When is the exponential equation preferable? When is the hydrostatic equation preferable? In questions two and three above, which equation did you use, and would your answers change if you used the other instead? Think about the assumptions that go into each equation and which variables are assumed to remain constant while pressure varies.

Selected Practice Question Answers:

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CHAPTER OVERVIEW

2: Solar and Infrared Radiation

Alison Nugent

Learning Objectives

By the end of this chapter, you should be able to:

- Define black body radiation and Planck's Law
- Apply Wien's law to compute the maximum emission wavelength
- Use Stefan-Boltzmann's law to compute radiative emittance
- Describe Earth's surface radiation budget, including shortwave and longwave components
- Define and differentiate obliquity, eccentricity, and precession
- Describe the cause of seasons on Earth
- Describe the diurnal cycle of radiative fluxes

[2.1: Introduction](#)

[2.2: Radiation](#)

[2.3: Wave Propagation](#)

[2.4: Characterizing Emission](#)

[2.5: Application to the Earth-Sun System](#)

[2.5.1: Albedo](#)

[2.5.2: Surface Energy Balance](#)

[2.5.3: Radiation Interaction with the Atmosphere](#)

[2.5.4: Radiation Changes with Time](#)

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2.1: Introduction

Outside on a sunny day, you can feel the sun's energy on your skin. You can feel it in the form of *heat* due to the transfer of energy between objects. Standing under the shade of a tree (which blocks the sun's rays) makes a significant difference in temperature and your probability of getting a sunburn. The closest Earth gets to the Sun is approximately 93 million miles. How does the sun's energy reach so far? The answer is in radiation. Radiation is the primary mechanism of energy transfer on Earth, including the transfer of energy from the Sun to the Earth over great distances through the vacuum of space.

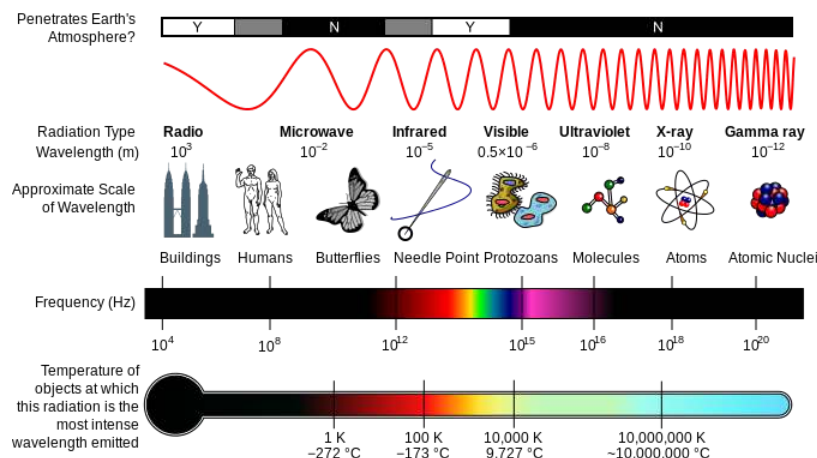


A setting sun with an orange sky shining over a breaking ocean wave ([CC BY 2.0](#)).

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2.2: Radiation

Radiation can be thought of in two ways: electromagnetic waves or as photons. For the purpose of atmospheric science, we will generally consider radiation as a wave rather than a photon particle. **Electromagnetic radiation** is a type of energy produced by electric and magnetic fields, taking a variety of names depending on the wavelength. For example, you've probably heard of radio waves and x-rays, whereby the primary difference is the wavelength of the waves in each. The next image shows the relationship between wavelength, frequency, and temperature. We will go into further detail later, but colder objects radiate at lower frequencies and longer wavelengths, and warmer objects radiate at higher frequency and shorter wavelengths. This relationship holds true across all scales, from atomic nuclei to planets, and across temperatures from near absolute zero to 10s of millions of Kelvins.



The electromagnetic energy spectrum, including penetration of Earth's atmosphere (top), wavelength visual (red line), radiation type (name given), wavelength value (m), wavelength scale (image), frequency (Hz), and temperature of black body objects that emit at that wavelength of radiation ([CC BY-SA 3.0](#)).

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2.3: Wave Propagation

How fast does the Sun's electromagnetic radiation travel to Earth and how can we characterize the waves? All electromagnetic waves travel at the *speed of light* (often given the variable " c_0 ", approximately $3 \times 10^8 \text{ m}\cdot\text{s}^{-1}$). The waves have a wavelength (λ) given by the distance from one wave crest to the next. The waves also have a frequency (ν), which is the number of repeated wave occurrences in a specified period of time. The unit for frequency is Hertz (cycles per second). Finally, one can also define a wavenumber for waves, which is one over the wavelength, or the number of waves in each meter, $\sigma \text{ (cycles}\cdot\text{m}^{-1}) = 1/\lambda$. Circular frequency is defined as $\omega \text{ (radians}\cdot\text{s}^{-1}) = 2\pi \cdot \nu$.

$$\text{Wavelength} \times \text{Frequency} = \text{Speed of light}$$

$$\lambda \times \nu = c_0$$

With variables for wavelength and frequency, we can very clearly define any type of wave.

Hawaii Focus Box

Electromagnetic waves are simply a type of wave. They can be described with some of the same words and variables as ocean waves. If you are a surfer, you probably check the swell forecast for wave height and wave period before you head out to the water. The best waves for surfing have a long period, which helps to separate and clearly define individual waves. Wave periods larger than 12 or 14 seconds are ideal. Depending on whether you are a beginner or advanced, you might like to have a wave face height between 2 and 30 feet, respectively.



A breaking ocean wave ([NOAA Public Domain](#)).

Wave period is defined as the time it takes for a wave to complete one cycle. Wave period is simply the inverse of wave frequency.

Wave face height is related to both wave amplitude and frequency, but is much more complicated because in the case of surfing, it is measured as the wave breaks. In the case of electromagnetic energy, the waves do not break.

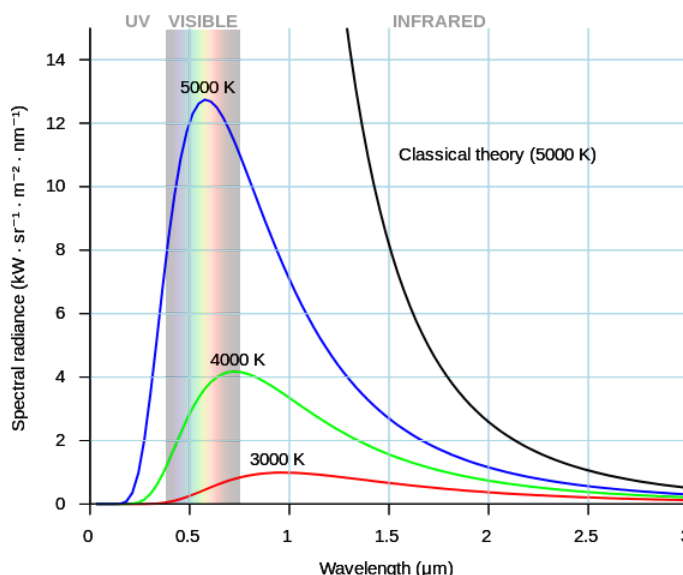
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2.4: Characterizing Emission

Where does this electromagnetic energy come from? It surrounds us every moment of every day in many forms. In fact, any object warmer than absolute zero (0 K) emits radiant energy. In order to estimate the amount of radiant energy an object emits, a common simplification is needed: we assume an object behaves as a blackbody. A **blackbody** is an object that emits and absorbs the maximum amount of radiation possible given its temperature.

Planck's Curves

Planck's curves are used to show the amount of emitted radiation and primary wavelengths of electromagnetic energy that a black body emits given its temperature. The diagram below shows multiple Planck function curves for various temperature black bodies. The red line denotes an object that has a temperature of 3000 K, the green line is 4000 K, and the blue line is 5000 K. As the black body becomes hotter, it emits at shorter wavelengths and greater intensities.



The spectral radiance (y-axis, emitted radiation) vs. the wavelength (x-axis) of radiation emitted from black bodies of various temperatures. As the temperature of the black body increases, the amount of emitted radiation increases, and the radiation is emitted at shorter and shorter wavelengths ([Public Domain](#)).

Wien's Law

Planck's function tells us the amount of emitted radiation and wavelengths over which it is emitted given the temperature of the black body. We can use another law to determine the maximum wavelength emitted by a black body. **Wien's Law** states that the shorter the wavelength emitted, the hotter (more kinetic energy) the object is. In Wien's equation, sometimes the numerator is given as "a", a constant equal to 2897.

$$\text{Maximum Wavelength} = \frac{a}{\text{Temperature}} = \frac{2897}{T}$$

$$\lambda_{\text{max}} = \frac{2897}{T}$$

Using Wein's equation to find wavelength gives an answer in microns, μm. One micron is equal to 10⁻⁶ meters.

✓ Example 2.4.1: Solar Radiation Wavelength

Our sun is a yellow dwarf star with a surface temperature near 5800 K. What is the maximum wavelength of the radiation it emits? Where does that radiation fall on the electromagnetic spectrum?

Solution

Wein's Law gives us an equation that relates temperature to the maximum wavelength of radiation emitted by an object:

$$\lambda_{\max} = \frac{2897}{T}$$
$$\lambda_{\max} = \frac{2897}{5800K} = 0.499$$

This radiation, 0.499 μm , is in the range of visible light, given as $0.5 \times 10^{-6} \text{ m}$ by the chart earlier this chapter.

Stefan-Boltzmann Law

One last very important law for radiation will be discussed here. The **Stefan-Boltzmann Law** relates the total radiation emitted (total emitted power per area) to the area under Planck's curve. This can be used to show that the hotter the object, the more energy it radiates per unit area.

$$\text{Emission} = \frac{\text{Power}}{\text{Area}} = \text{Stefan - Boltzmann Constant} * \text{Temperature}^4$$

$$E = \sigma * T^4$$

In the above equation the amount of radiance emitted per area is equal to the temperature of the black body raised to the 4th power. This relationship is extremely important. It shows that the amount of radiation emitted depends heavily on temperature such that small temperature fluctuations result in large changes in emittance.

The Stefan-Boltzmann constant is $5.67 \times 10^{-8} \text{ W} \cdot \text{m}^{-2} \cdot \text{K}^{-4}$ and the symbol sigma, σ , is used.

Pro Tip: Whenever you see the Stefan-Boltzmann constant being used (σ), you know that the assumption of a black body has been made.

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2.5: Application to the Earth-Sun System

In the prior section, the discussion turned rather technical. We went down a rabbit hole of radiation wavelengths, total emitted power per area, and the assumption of black bodies. We did this so that we could understand basic relationships about the energy balance in the Earth-Sun system. The Sun's average temperature is above 5,000 K while the Earth's average temperature is in the range 210-310 K (we will discuss this further in a later chapter). This means that the Sun and Earth radiate energy very differently.

The Sun emits solar radiation, also known as *ultraviolet radiation* or *shortwave radiation*. The Earth emits *infrared radiation* or *longwave radiation*. This follows directly from the electromagnetic energy spectrum and the respective temperatures of the Sun and Earth. The Sun emits radiation at a shorter wavelength than the Earth because it has a higher temperature, and Planck's curve for higher temperatures peaks at shorter wavelengths. It is for this reason that Earth's radiation is referred to as longwave, and the Sun's radiation is called shortwave.

We learned that a black body absorbs all incoming radiation and emits the maximum possible radiation given its temperature. In the Earth-Sun system, this means that the Earth absorbs all incoming radiation from the Sun, and emits the maximum amount given its temperature. In practice it is a little bit more complicated than this.

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2.5.1: Albedo

Not all incoming solar radiation is absorbed by the Earth because the Earth is not a perfect black body. Instead of absorbing all incoming radiation, some of it is reflected. Reflection refers to radiative energy bouncing back away from an object. We can define **albedo** (α) as the ratio of the amount of radiation reflected from an object to the amount of radiation received by an object.

$$\text{Albedo} = \alpha = \frac{\text{Reflected}}{\text{Incident}}$$

Pro Tip: The simplest way to think of albedo is based on the color of the object. Objects that are white colored are highly reflective and have a high albedo because the amount of reflected light is large compared to the amount of incoming incident light. You'll know this if you've been on a white sandy beach or a snow field and felt the brightness of the environment. Following the same logic, dark surfaces absorb light so black, brown, or dark green surfaces typically have a low albedo. You'll know this if you've walked across a black paved surface and felt the temperature difference between the blacktop and the white painted parking space demarcations.

Earth has an albedo of about 0.3 or 30%. This is an average that takes into account the high albedo and high latitude regions that are snow covered as well as the clouds and the much lower albedo oceans. Note that radiation reflected from an object does not warm the object.

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2.5.2: Surface Energy Balance

Arguably the most important aspect to consider about the Earth-Sun system is the energy balance. In **steady-state**, the amount of incoming energy should equal the amount of outgoing energy (Net Radiative Flux = $F^* = 0$).

Let's start with the incoming solar radiation. The *solar constant* " S " is approximately equal to $1361 \text{ W}\cdot\text{m}^{-2}$. This value is a rough estimate of the amount of energy per area received by the Earth from the Sun, but it is not exact. We call it a solar "constant" but it can be slightly lower or higher at times. Radiation emitted from a spherical source, like the sun, decreases by the square of the distance from the sphere's center. This is called the inverse square law.

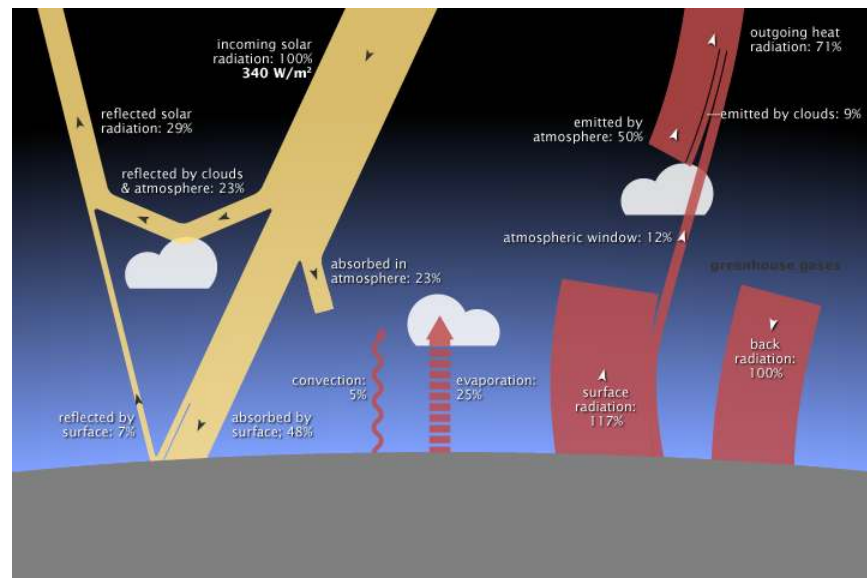
The following equation is a basic budget equation. Net radiation (F^*) is equal to the incoming solar radiation ($K \downarrow$), the reflected solar radiation ($K \uparrow$), the longwave radiation emitted by the Earth ($I \uparrow$), and the downwelling longwave radiation emitted from the atmosphere ($I \downarrow$) received by the Earth's surface.

$$F^* = K \uparrow + K \downarrow + I \uparrow + I \downarrow$$

Some of the sunlight that reaches the surface of the Earth is reflected, depending on the albedo, $K \uparrow$ is often written in the following way:

$$K \uparrow = -\alpha \times K \downarrow$$

This is a brief introduction to a surface energy balance model. As you may imagine, it can become much more complicated depending on the factors involved. It also strongly depends on the number of layers considered in the model. We will discuss this further in a later chapter.



Earth's radiative balance, incoming and outgoing radiation. ([NASA Public Domain](#)). Description in the text below.

For now, you should understand that incoming solar radiation is called shortwave radiation and is in the ultraviolet and visible portions of the electromagnetic spectrum because of the emission temperature of the Sun. When solar radiation interacts with the Earth, it is partially absorbed by the Earth's surface, and partially reflected, depending on the albedo of the surface. In the diagram above, you can see that some of the incoming solar radiation is reflected by clouds, some is reflected by the Earth's surface, but most is absorbed by the Earth's surface or the atmosphere.

You should also understand that Earth emits radiation too. However, it is at a lower intensity and a much longer wavelength, which is called the infrared portion of the electromagnetic spectrum because of the lower emission temperature of the Earth. Radiation is emitted by the Earth's surface, and by the atmosphere. We'll go into more detail on this later.

Pro Tip: Many synonyms were discussed above. When referring to radiation from the sun you may hear the following words used: solar radiation, shortwave radiation, and ultraviolet radiation. When referring to radiation from the Earth you may hear

the following words used: infrared radiation, longwave radiation.

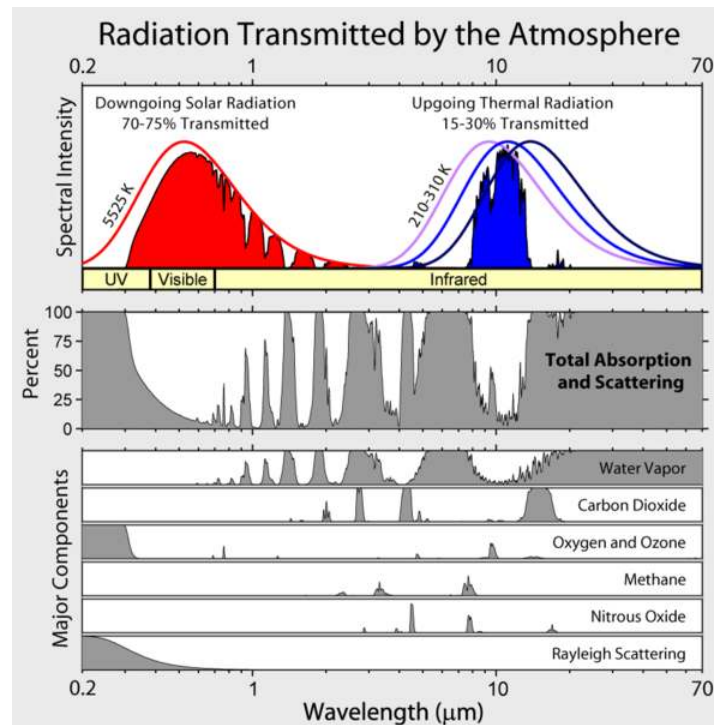
Earth's average temperature remains relatively constant as there is a balance of outgoing radiation and incoming radiation.

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2.5.3: Radiation Interaction with the Atmosphere

The diagram below shows the spectral intensity of downgoing solar radiation (UV and visible in red) and upgoing thermal radiation (IR in blue). The panels below the spectral intensity graph show the total percentage of radiation absorbed and scattered by the atmosphere as a function of wavelength, and divided up by the primary greenhouse gases in the atmosphere. **Greenhouse gases** are gases in the Earth's atmosphere that can absorb and emit radiation, primarily infrared radiation.

The greenhouse gas that interacts with radiation across the largest number of wavelengths is water vapor (H_2O), which interacts strongly with the atmosphere, especially in the longer wavelength portion of the spectrum. Other gases, such as Carbon Dioxide (CO_2), Oxygen (O_2), Ozone (O_3), Methane (CH_4), and Nitrous Oxide (NO) also interact with radiation.



Radiation transmitted by the atmosphere as a function of wavelength. Downgoing solar radiation is on the left, and upgoing thermal radiation is on the right, with the percentage of radiation absorbed and scattered by the Earth's atmosphere below. Finally, the percentage of radiation absorbed and scattered is then divided by the major greenhouse gases in the panels at the bottom ([CC BY-SA 3.0](#)).

In the panel of the figure above showing the total percentage of radiation that is absorbed by the Earth's atmosphere, you'll notice white gaps. These are called **Atmospheric Windows** where the wavelength of radiation is able to pass through the atmosphere without interaction with greenhouse gases. Particularly notable is the visible portion of the electromagnetic energy spectrum that can pass through the atmosphere unimpeded. Also notable is the large amount of interference in the infrared portion of the spectrum. Earth's atmosphere is practically transparent for shortwave radiation, and it strongly absorbs infrared radiation. This will be important for climate factors discussed in a later chapter.

Open-ended Question: Why do you think visible light (the peak of down-going solar radiation) is centered on an atmospheric window?

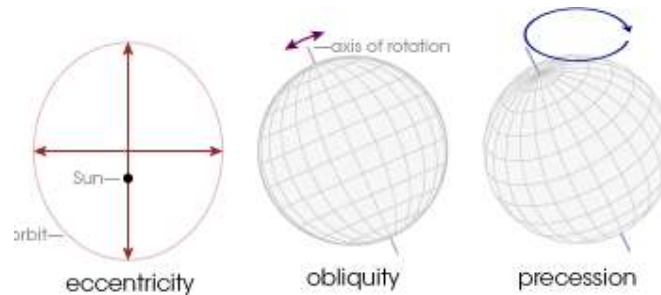
The interaction of atmospheric gases in Earth's atmosphere is an important part of its energy balance. The *greenhouse effect* is the result of greenhouse gases absorbing and emitting outgoing infrared radiation emitted from Earth's surface. Because Earth's outgoing longwave radiation is partially absorbed by the atmosphere, this has a warming effect on Earth's surface making it warmer than it would be otherwise. While the greenhouse effect has a bad reputation, this process has in fact allowed Earth to be habitable.

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2.5.4: Radiation Changes with Time

The amount of radiation the Earth receives from the Sun varies over many time scales, from thousands of years, to one year, to daily time periods. These will be discussed in the following sections.

Changes in solar radiation on both daily and annual time scales can be explained by Earth's orbital path around the Sun. These aspects will be discussed more thoroughly in a later chapter, but for now, let's have a brief introduction. There are three primary factors to consider: eccentricity, obliquity, and precession. **Eccentricity** is the circularity of a planetary orbit. For example, a circle has zero eccentricity. **Obliquity** is the degree of tilt in the axis of rotation. Finally, **precession** is the wobble in the rotational axis of a planet that slowly traces out a cone. The three orbital parameters are shown in the image below.

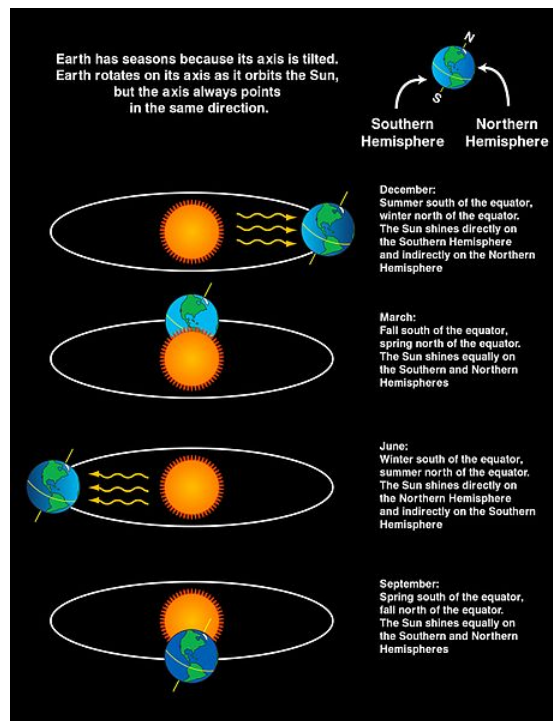


Visual displaying three orbital parameters: eccentricity, obliquity, and precession ([NASA Public Domain](#)).

Currently the Earth has a 0.0167 orbital eccentricity, 23.44 degree tilt from vertical, and precesses on very long time scales. In fact, all of these factors change slightly over very long time scales but for now, let's consider them to be constant.

Seasonal Changes

Seasons occur due to changes in solar radiation that come from the position of Earth with respect to the Sun. The diagram below shows the position of the Earth with respect to the sun during each of the four seasons.



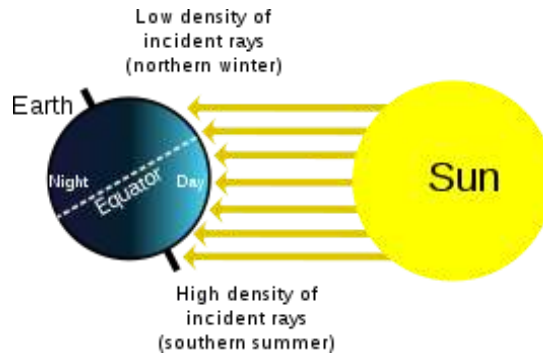
Schematic diagram showing the reason Earth has seasons ([CC BY-SA 4.0](#)).

During summertime in each hemisphere, the hemisphere is facing toward the sun. For example, in June in the Northern Hemisphere summer, the sun shines more directly on the Northern Hemisphere than the Southern Hemisphere. While it seems like a small change, the sun angle and the amount of solar radiation absorbed (per area) varies significantly throughout the year. This is because

in addition to changing the angle of the Sun, the position of the Earth also changes the length of day throughout the year. Seasons are due to the tilt of the Earth. If the Earth's rotation was not tilted, it would not have seasonal changes.

Pro Tip: The cause of seasons on Earth is the topic of a common misperception. Earth's seasons are caused by its tilted orbit. While it is true that Earth's orbit is not spherical, and it is closer to the Sun during Northern Hemisphere winter, the distance from Earth to the Sun is not the cause of seasons.

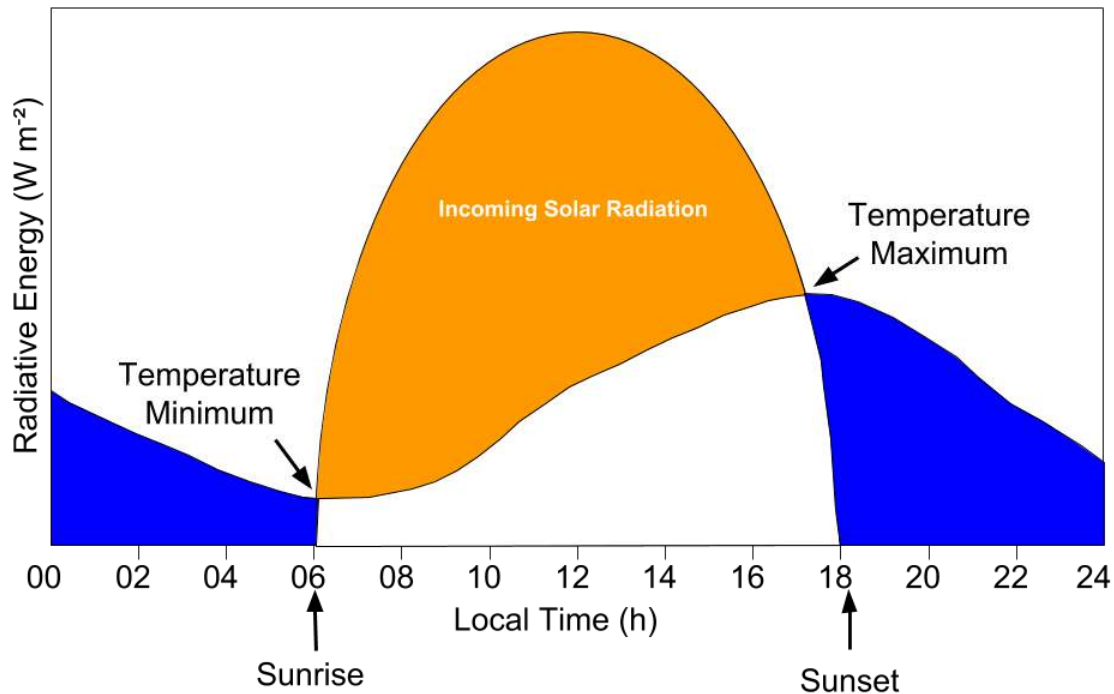
The diagram below shows the summer season in the Southern Hemisphere where there is a higher density of incident rays due to the higher Sun angle. The Northern Hemisphere is experiencing wintertime. When the Sun's rays are at an angle as they are in the Northern Hemisphere, the same amount of energy is spread over a larger area than they would be if the Sun's rays were perpendicular to the surface. Again, the angle of the Sun's rays and the length of day change because of the Earth's tilt.



Distribution of solar rays on Earth, with the summer season receiving the majority of the solar radiation ([Public Domain](#)).

Daily Changes

Daily changes are also called “diurnal.” The Earth absorbs radiation from the Sun during the daytime. This is only true for one location, but accounts for the increase in temperature throughout the day from a point perspective. What we likely have not experienced directly is that the Earth emits infrared radiation all day and night. The lack of incoming solar radiation and the emission of infrared radiation is what accounts for the decrease in temperature at night. The combination of nonstop outgoing longwave radiational cooling from Earth's radiative emissions and daytime solar shortwave radiational heating results in a diurnal cycle of net radiation and temperature, as seen below.



The amount of radiative energy (y-axis, in W m^{-2}) as a function of time (x-axis, in hours). Yellow shows net warming and blue shows net cooling. The temperature is coolest at sunrise, warms during the day and reaches its maximum temperature in the late afternoon. (Image Created by Britt Seifert).

In this chapter we focused on the radiation: how we can define and describe it by wavelength and intensity; the temperatures it corresponds to as well as how it changes over time scales. As a reminder, these were our learning goals:

1. Define black body radiation and Planck's Law
2. Apply Wien's law to compute the maximum emission wavelength
3. Use Stefan-Boltzmann's law to compute radiative emittance
4. Describe Earth's surface radiation budget, including shortwave and longwave components
5. Define and differentiate obliquity, eccentricity, and precession
6. Describe the cause of seasons on Earth
7. Describe the diurnal cycle of radiative fluxes

In the following chapter, we'll begin to see how this radiation does work, and what that means for the environment.

Chapter 2: Questions to Consider

1. Are the following statements about radiation true or false?
2. Describe the relationship between wavelength, energy emission, and the temperature of an object.
3. If the temperature of the Earth is 257 K, what is the total radiative flux emitted? What is the peak emission wavelength?
4. How would the seasons change if the Earth's tilt were greater?

Selected Practice Question Answers:

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CHAPTER OVERVIEW

3: Thermodynamics

Alison Nugent

Learning Objectives

By the end of this chapter, you should be able to:

- Define and describe four methods of energy transfer
- Describe the change in energy associated with changes in water state
- Define and apply the first law of thermodynamics
- Differentiate Eulerian and Lagrangian frameworks
- Describe the importance of the dry adiabatic lapse rate, and recall what sets its constant value in the atmosphere
- Compute potential temperature and apply the conserved variable approach
- Draw a diagram of surface heat fluxes and Earth's radiation budget
- Compute the Bowen ratio, and define latent and sensible heat flux

[3.1: Introduction](#)

[3.2: Energy Transfer](#)

[3.3: Latent Heat](#)

[3.4: Specific Heat](#)

[3.5: First Law of Thermodynamics](#)

[3.6: Frameworks for Understanding the Atmosphere](#)

[3.7: Introduction to Thermodynamic Diagrams](#)

[3.8: Additional Information](#)

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3.1: Introduction

What is Energy? You may hear frequently about “green energy”, “clean energy”, “renewable energy”, and “solar energy” in the media, as energy is currently a hot topic. Power plants, wind turbines, and solar panels may come to mind. The first paragraph in Roland Stull’s *Practical Meteorology* Chapter 3 discusses several types of energy right off the bat, but first of all, what is energy?



Convective clouds in the atmosphere are driven by an enormous source of energy called latent heat ([CC BY-SA 2.0](#)).

Your intuition will probably tell you that we need energy in order to do things. Energy makes stuff “go”. Your appliances, your car, and your body all need energy. Your body utilizes energy to maintain its various functions even as you read this sentence. You know that energy is a real thing that exists; you see evidence of it everywhere. The heat from your stovetop, the ice melting in your glass, and that thunderstorm rolling in are all evidence of energy at work, on scales both micro and macro.

Energy can be defined as the ability to do **work**. When you apply a force on an object, it is said that work is done on the object if that object is displaced, meaning it moves from its original location. For example, when you pick up a book, you exert a force against gravity causing the book to change position, and you do work on the book. The higher you lift the book or the further you throw it (should you decide to), the more work you do. However, if you apply a lot of force on a heavy piece of furniture but it doesn’t move from its original place, no work has been done regardless of the amount of effort.

$$\text{Work} = \text{Force} \times \text{displacement}$$

$$W = F \times \Delta d$$

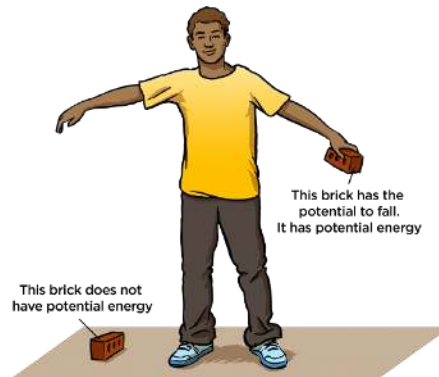
Note

Δ is the Greek letter delta, and denotes a change. Here, Δd is the change in position, or the displacement. The standard unit of force is a Newton ($N = \text{kg} \cdot \text{m} \cdot \text{s}^{-2}$), which is defined as the force required to make a mass (m) of 1 kg accelerate at $1 \text{ m} \cdot \text{s}^{-2}$. Force (F) is equal to the mass (m) of an object times its acceleration (a): $F = m \times a$.

Internal energy is the total amount of energy stored in any object and determines how much work the object is capable of performing. This includes both **kinetic energy** (energy that an object has when it is in motion) and **potential energy** (energy that is stored). For example, a bowling ball sitting on a table contains energy despite the fact it is not in motion. It does not have kinetic energy because it is still, but it contains potential energy simply because of where it is situated. Were it nudged off the table, the bowling ball will do work because it will be pulled downward by gravity. This is an example of *gravitational potential energy*. The potential energy (PE) due to gravitational pull is given by the following equation:

Potential energy = mass $\times$ gravitational acceleration $\times$ height above ground

$$PE = m \times g \times h$$



Gravitational potential energy ([CC BY 2.0](#)).

Anything that moves contains kinetic energy (KE), which is given by the following equation

$$KE = \frac{1}{2}m \times v^2$$

where m is the mass of an object in kilograms (kg) and v is the velocity of an object in meters per second ($\text{m}\cdot\text{s}^{-1}$). From this relationship you can see that objects with more mass or objects that are moving faster have more energy.

Energy takes on many forms and often changes forms from one to the other, but the total amount of energy in the universe remains constant. Energy cannot be created or destroyed. This means that the energy lost in a process must be the same as the energy gained in another. This is what the law of conservation of energy means, and this is what is known as the **first law of thermodynamics**. The first law of thermodynamics frequently comes into play in atmospheric motions and will be discussed further later in this chapter.

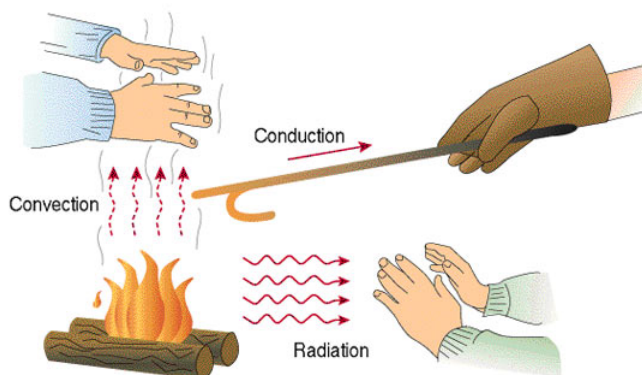
In short, energy is the capacity of a system to perform work. Energy is always conserved and cannot be created or destroyed. We begin this chapter with a short review of energy because energy is ultimately responsible for Earth's weather from temperature changes in the atmosphere to the resulting air motions. Without energy, no weather would occur.

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3.2: Energy Transfer

One important example of kinetic energy is **thermal energy**, which comes from the tiny movement of many molecules in a system. In Chapter 1, we discussed how *temperature* is a measure of the average speed of atoms and molecules in a system. Here we can further describe temperature as being proportional to the average kinetic energy of the random motions of the molecules in a system. The faster the molecules move, the higher the temperature.

The transfer of thermal energy due to the temperature difference between two objects is what is known as **heat**. Heat is a form of energy in transit, and once transferred, it is stored as internal energy. There are four main methods of heat transfer: *conduction*, *convection*, *radiation*, and the absorption or release of *latent heat*.

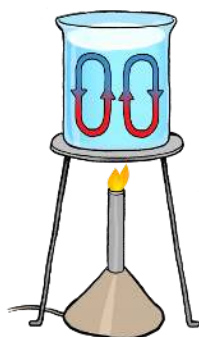


Conduction, convection, and radiation are all methods of heat transfer ([CC BY-SA 4.0](#)).

When you directly touch a hot object, such as the stovetop, the energy from the hot stove top is immediately transferred to your cool hand due to a difference in the speed of the molecules, causing you to feel a burn. This is an example of **conduction**: energy directly transferred through a substance without the movement of material. Certain materials are better conductors of heat than others. Metal, for example, conducts heat very efficiently, while air, which acts as an insulator, is a very poor conductor of heat.

Another type of energy transfer is convection, which occurs in liquids and gases (both fluids) because molecules can move freely and currents can naturally occur in them. This happens constantly in the atmosphere. **Convection** refers to movement within a fluid due to the tendency of lower density fluid to rise over higher density fluid, which sinks due to the force of gravity resulting in heat transfer within the fluid.

In the figure below, a beaker is being heated from the bottom by a flame. The arrows show upward movement in the center where the fluid is heated and therefore is less dense and buoyant. The cooler fluid at the top of the beaker is more dense and sinks toward the bottom under the influence of gravity.



Arrows show the motion of convection within a fluid that is heated from the bottom ([CC BY 2.0](#)).

Finally, when it comes to the atmosphere, the most important form of energy is the energy we get from the Sun, which is called *radiant energy* or radiation. **Radiation** is another type of heat transfer that was covered in Chapter 2. Remember that radiant energy can be transferred to an object without the space in between being heated. This is the result of electromagnetic waves from the Sun,

which are then absorbed by the Earth's surface or atmosphere and converted to thermal energy. Electromagnetic waves do not need any matter to travel through and propagate at the speed of light in a vacuum which is about $300,000 \text{ km}\cdot\text{s}^{-1}$.

Heat, or the transfer of thermal energy due to the difference in temperature between two objects, is defined as Δq and is given in standard units of joules per kilogram ($\text{J}\cdot\text{kg}^{-1}$). One joule (J) is the standard unit of energy or work ($\text{kg}\cdot\text{m}^2\cdot\text{s}^{-2}$).

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3.3: Latent Heat

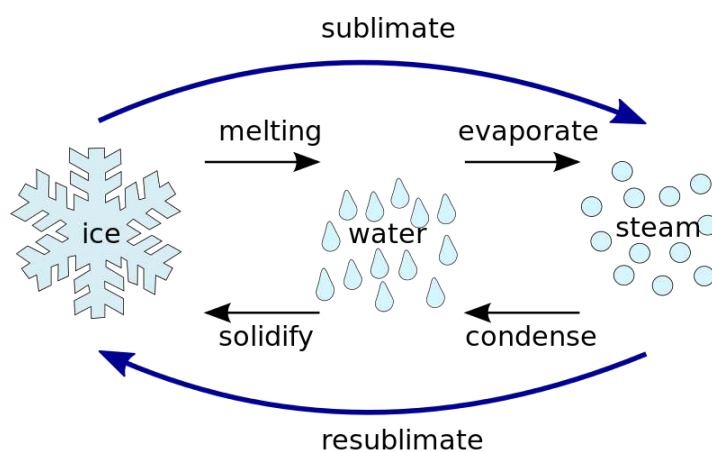
Latent heat refers to the amount of heat that is added or removed from a system when a change of phase takes place. In atmospheric science, this almost always refers to the phase changes of water (H_2O).

When heat is added to an ice cube, the temperature of the ice will increase until it reaches its melting point ($0^\circ C$). After this point, any further addition of heat will cause the ice to melt to liquid water, but the temperature of the ice-water system will not change until the phase change is complete. The heat that is absorbed by the ice-water system from the environment in order for the phase change to occur is known as latent heat.

You have experienced this as your ice melts in your water glass. It stays cold until all of the ice is gone and then rapidly warms. The absorption of latent heat is also the reason you feel cold right after leaving your shower or a body of water. The excess water on your skin starts to evaporate, but it requires additional energy in order to transition from liquid to vapor. Energy is absorbed from the environment (your skin) to evaporate the remaining liquid, causing your skin to cool. This is known as evaporative cooling.

Processes that take heat from the environment (melting, evaporation, sublimation) are considered cooling processes. On the other hand, processes that add heat to the environment (condensation, freezing, deposition) are warming processes because during the phase change energy is released. This is an important distinction, so be sure this is clear before moving on. I'll repeat it below for emphasis.

Pro Tip: Melting, evaporation, and sublimation are considered cooling processes because they take heat from the environment. Condensation, freezing, and deposition are considered warming processes because during the phase change energy is released and they add heat to the environment.



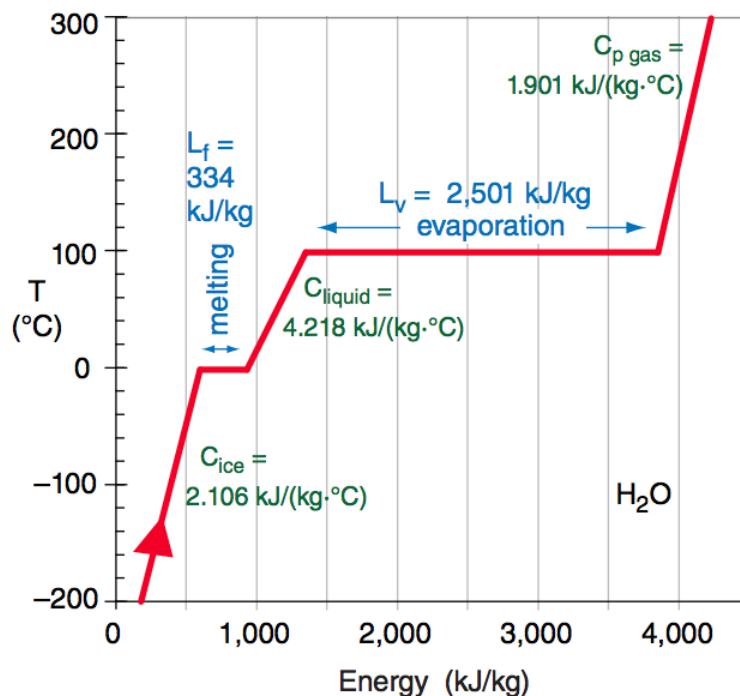
Phase changes of water ([CC BY-SA 3.0](#)). The lower energy state is on the left, and the higher energy state is on the right. Going from left to right takes energy (cools the environment), going from right to left releases energy (warms the environment).

The reason phase changes of water are so important in the atmosphere is because it acts as the energy to fuel convection and thunderstorms. When water vapor condenses to form the liquid water found in clouds, this process releases enormous amounts of latent heat to the environment.

Latent heat (Q_E) is the latent energy that is possessed by an object of mass m , but usually we want to know the change of latent heat that is caused by a phase change process.

$$\Delta Q_E = L * \Delta m_{\text{water}}$$

The change in latent heat is equal to the latent heat factor multiplied by the mass of water undergoing a phase change.



The y-axis of this figure shows the state of water at different temperatures. The x-axis shows the amount of energy applied to the water. When water reaches 0°C and 100°C , the melting/freezing and boiling points of water, respectively, the temperature stops increasing while energy is still being applied. Instead of increasing the water's temperature, this energy is instead going into the processes of melting and evaporation. From here, you can see that the latent heat of vaporization is much greater than the latent heat of melting, because much more energy is applied while the temperature remains the same ([CC BY-NC-SA 4.0](#)).

Different phases of water have different latent heat factors (L), meaning the change in phase of a mass of water corresponding to the change in latent heat is different depending on what process is taking place. This is well displayed in the image above by the horizontally flat regions where energy is going into the system but no temperature change is occurring.

For evaporation or condensation (phases changes between liquid water and vapor), the latent heat of vaporization, $L_v = 2.5 \times 10^6 \text{ J}\cdot\text{kg}^{-1}$ is used. For melting or freezing (phase changes between ice and liquid water), the latent heat of fusion or freezing, $L_f = 3.34 \times 10^5 \text{ J}\cdot\text{kg}^{-1}$ is used.

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3.4: Specific Heat

Latent heat is the term for the amount of heat released from changes in phase, but what about changes in temperature unassociated with phase changes? These portions are the slanted lines in the figure above where energy is going into the system and the temperature is responding. **Specific heat or heat capacity** (C) is the term for the amount of heat required to raise the temperature of a substance by a certain amount. Typically this is defined at a constant pressure (C_p) or a constant volume (C_v).

Some substances, such as water, require a particularly large amount of heat energy in order for the temperature to change. The heat capacity of a substance is the ratio of heat energy that is absorbed to the corresponding rise in temperature. The specific heat is the heat capacity of a substance per a unit mass. Basically, specific heat is the amount of heat that is needed in order to raise the temperature of one gram (g) of a substance by one degree Celsius, or Kelvin.

For example, the specific heat of pure water is 4186 Joules per degree Kelvin per kilogram ($\text{J}\cdot\text{K}^{-1}\cdot\text{kg}^{-1}$), while the specific heat of dry air at sea level (C_{pd}) is about $1004 \text{ J}\cdot\text{K}^{-1}\cdot\text{kg}^{-1}$. The specific heat of dry air at constant volume (C_{vd}) is equal to $717 \text{ J}\cdot\text{kg}^{-1}\cdot\text{K}^{-1}$. This means that it takes much more energy to raise 1 kg of water by 1°C than it does for 1 kg of dry air. Remember that Celsius and Kelvin have a linear relationship so the change in temperature of one degree Celsius is the same as a change of one degree Kelvin ($K = ^\circ\text{C} + 273$).

Hawaii Focus Box

The difference in specific heat capacity of water and land leads to different annual cycles of air temperature and water temperature in Hawaii. The heat capacity of sea water is about $3985 \text{ J}\cdot\text{K}^{-1}\cdot\text{kg}^{-1}$ while the heat capacity of land is typically less than $1000 \text{ J}\cdot\text{K}^{-1}\cdot\text{kg}^{-1}$. This means that land heats up faster and cools down faster while the ocean takes longer to warm and also takes longer to cool.

The warmest temperatures in Hawaii typically follow the solar cycle with a slight lag. The warmest temperatures in the Northern Hemisphere occur during the time periods where the Sun's angle is highest and the length of day is longest as we discussed in Chapter 2. This typically corresponds to June through August. However, the warmest ocean temperatures lag the land temperatures significantly by a few months because water takes more energy to warm. The warmest ocean temperatures occur from August through October.

There is some feedback where the warm ocean temperatures affect land temperatures as well, but it all comes down to the difference in specific heat.

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3.5: First Law of Thermodynamics

As stated previously, the first law of thermodynamics states that energy is neither created nor destroyed. The total amount of energy must be conserved. In the atmosphere, this means that the amount of heat applied to a mass of air (thermal energy input) must equal the total sum of the warming of the air, plus the amount of work done per unit mass of air.

When heat (Δq , $\text{J}\cdot\text{kg}^{-1}$) is added to a mass of air (often referred to as an air parcel, see below), some of the added thermal energy warms the air, which increases its internal energy. When the air parcel warms, one of two things must happen. Recall the ideal gas law from Chapter 1: $P = \rho \times R_d \times T$. If temperature increases, (1) density must decrease to keep pressure constant or (2) the pressure will increase. Small pressure differences in the atmosphere always equalize first, so as the air warms, the density of the air parcel decreases. The decreased density of the air parcel gives it a lower density than the surrounding air, and it is therefore buoyant and begins to rise. As the parcel rises, it expands in volume in order to maintain equilibrium with the lower pressure outside the air mass, and it pushes into the surrounding atmosphere. Because of this, some of the thermal energy that is added to the air goes into doing the work of expansion and not all of it is used for warming.

$$\text{Heat Added} = \text{Warming} + \text{Work done by the expanding air}$$

$$\Delta q = C_v \times \Delta T + P \times \frac{\Delta V}{m}$$

For the atmosphere, a more usable form of the First Law of Thermodynamics is:

$$\text{Heat Transferred} = \text{Enthalpy Change} - \text{Pressure Change per Mass}$$

$$\Delta q = C_p \times \Delta T - \frac{\Delta P}{\rho}$$

In the atmospheric form of the First Law of Thermodynamics, $C_p \times \Delta T$ can also be redefined as Enthalpy (h), a term to quantify the total heat content of an air parcel.

$$\text{Enthalpy Change} = \text{Specific Heat (@ const } P) \times \text{Temperature Change}$$

$$h = C_p \times T$$

$$\Delta h = C_p \times \Delta T$$

This gives us the first term in the above equation for the first law of thermodynamics.

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3.6: Frameworks for Understanding the Atmosphere

All atmospheric science concepts are shared with other disciplines like physics and chemistry, but they often use a more specific set of equations or variables as shown above with the first law of thermodynamics. This helps to simplify things because general physics equations can be applied to nearly any form of matter, while in atmospheric science we primarily deal with gases that are not constrained to a specific volume.

In the two sections that follow, two important frameworks are described that will help you to view the atmosphere in ways that will simplify concepts in the sections and chapters to come.

Lagrangian vs. Eulerian

When we look at or try to solve a problem in the atmosphere, there are two different lenses or frameworks through which we can view the problem. An **Eulerian** framework is a fixed framework, relative to a single point on the Earth's surface. When a weather forecast is done for a given location on Earth or when you look at a dataset from one weather station, you are viewing the atmosphere from an Eulerian perspective — that is, how the wind and air travels past a fixed point. With a Eulerian framework, we need to be concerned about things like temperature and moisture advection, properties that travel with and are carried by the wind.

Another framework is **Lagrangian**, which is a framework that is constantly moving and travels with the air. When we are looking at motions within the atmosphere, such as rising or sinking air, it is useful to use this framework as a way to see how properties within the rising plume of air are changing. Both frameworks will be used in the coming sections and chapters.

Air Parcels

At this point, we should discuss a naming convention used in atmospheric sciences. Many times it is useful to think about a mass of air instead of individual molecules. We generally call a mass of air an “**air parcel**”. This is especially useful for distinguishing processes happening within the air, versus processes happening within the environment. An air parcel is often thought of as an amorphous bubble or blob of air, roughly the scale of a party balloon or a hot air balloon that contains uniform properties (temperature, density, pressure) throughout. Air parcels are simplified theoretical constructions used as a way to think about and examine motions and instability in the atmosphere.

For example, let's revisit the important topic of latent heating. We discussed how condensation of water vapor to liquid water is a warming process. This is only clear if we separate the processes happening within a parcel of air and within the rest of the environment. When water vapor condenses within an air parcel, it gives off energy, which acts to warm the environment. We will often refer to air parcels when we want to distinguish changes within a piece of air with respect to the rest of the atmosphere.

In reality, as air parcels move through the atmosphere, there will be some mixing between the air parcel and the environment, and heat can be exchanged with the environment or added via radiation. However, with the concept of air parcels, we simplify the situation and imagine that radiative effects are small, and that mixing only occurs on the outer edges of the parcel, with a protected inner core.

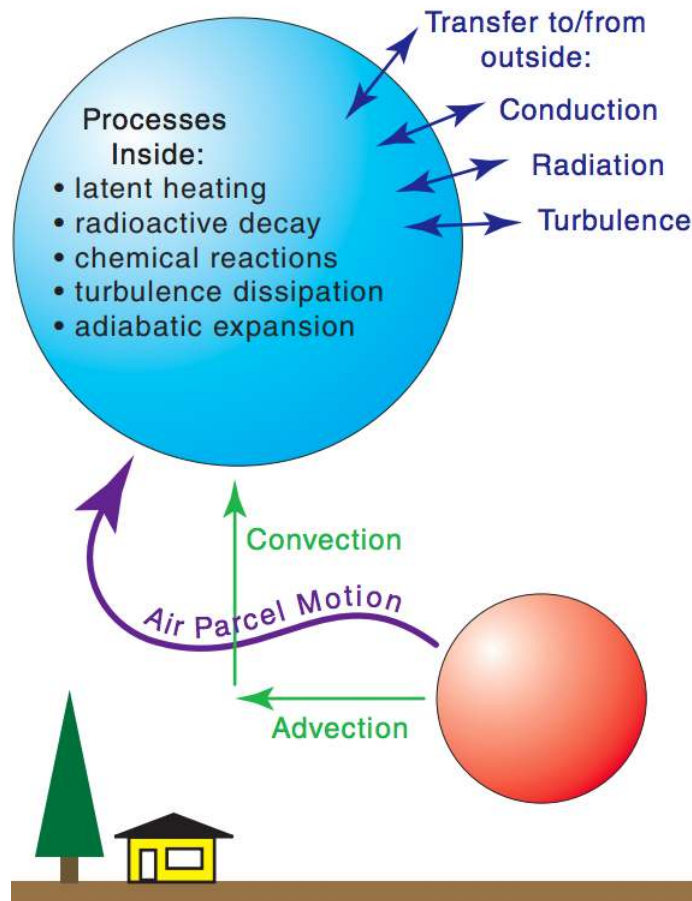
Defining Changes in the Atmosphere

In order to study changes in temperature, momentum, and moisture in an air parcel, a Lagrangian framework is always used. With a Lagrangian framework, we can see changes from the parcel's perspective as it moves, not from a fixed point on the ground.

Applying the hydrostatic equation from Chapter 1 to the first law of thermodynamics, we can get the Lagrangian first law of thermodynamics equation for a moving air parcel.

$$\Delta T = - \left(\frac{|g|}{C_p} \right) * \Delta z + \frac{\Delta q}{C_p}$$

Δq , or heat transfer, can be caused by various processes, outlined in the following figure. The figure shows an air parcel moving both horizontally (through advection) and vertically (through convection), as well as the various processes both inside and outside that are causing heat exchange. The temperature inside the air parcel is conserved unless heat is transferred to or from the environment, or if it loses or gains heat by rising or sinking, which will cause it to expand or contract, respectively.



The effects of surroundings on air parcel temperature and the corresponding heat exchange ([CC BY-NC-SA 4.0](https://creativecommons.org/licenses/by-nc-sa/4.0/)).

Lapse Rates

The atmospheric **lapse rate**, Γ , denoted by an upper-case gamma, is defined as the change in temperature with altitude, specifically *a reduction* in temperature with altitude.

$$\Gamma = -\frac{T_2 - T_1}{z_2 - z_1} = -\frac{\Delta T}{\Delta z}$$

A positive lapse rate indicates that the temperature is decreasing with increasing height, while a negative lapse rate indicates a “temperature inversion” meaning that the temperature is increasing with increasing height. Lapse rates are typically defined for:

1. The environment, Γ_e
2. A dry air parcel, Γ_d
3. A saturated air parcel (moist), Γ_m

We can observe the environmental lapse rate by using atmospheric sensors attached to weather balloons. As a weather balloon rises through the atmosphere, it measures temperature and other properties on the way. The environmental lapse rate varies depending on time of day, altitude, latitude, land surface properties, heat fluxes, and air movement. A typical Γ_e is around 6.5 K km^{-1} but this will be discussed in later chapters. You will have experienced this decrease in temperature with altitude if you have hiked in the mountains or seen an outdoor thermometer reading on a commercial flight.

Dry Adiabatic Lapse Rate

In atmospheric science, you will often hear the word “adiabatic”, typically accompanied by the words “cooling”, “warming”, or “lapse rate”. An Adiabatic process just means that there is no heat transfer taking place ($\Delta q = 0$) during the process. For an air parcel, this means that no thermal energy is entering or leaving the air parcel from the outside. However, internal processes are allowed, such as the ones shown in the figure above (most notably adiabatic expansion and latent heating). An adiabatic lapse rate

indicates that air is cooling or warming with altitude without any external heat exchange. An air parcel that contains no liquid water or ice (none of the moisture in the parcel has condensed into liquid, no saturation or latent heat release), will cool at the **dry adiabatic lapse rate**.

$$\Gamma_d = 9.8 \frac{K}{km} = 9.8 \frac{^{\circ}C}{km}$$

Pro Tip: A positive lapse rate means that the air temperature is decreasing with height.

The reason that an air parcel expands adiabatically as it rises is due to the fact that the environmental air pressure decreases with height. The air parcel's pressure will adjust to the lower pressure of its environment, but the parcel must expand in order to do so (in order for the air molecules inside the parcel to exert a smaller force). The parcel uses some of its own internal energy to do the work of expansion, and its temperature decreases as a result. The opposite is true as an air parcel sinks, the environmental pressure rises, and the parcel's pressure adjusts in order to maintain pressure equilibrium, and the parcel must shrink for its pressure to increase. As it shrinks, work is done on it, and the temperature rises.

In later chapters we'll define the moist or wet adiabatic lapse rate (Chapter 4), as well as the environmental lapse rate (Chapter 5) and their significance.

Potential Temperature

The **potential temperature**, θ , of a parcel completely ignores the temperature change of the parcel due to it having done work or been worked on (expanding or contracting). As a result, potential temperature is constant for an adiabatic process and θ does not change when $\Delta q = 0$. Potential temperature is proportional to the sensible heat contained in a parcel and can increase or decrease when sensible heat is added or removed through **diabatic** (non-adiabatic) processes ($\Delta q \neq 0$). Examples of diabatic heating include turbulent mixing, condensation (latent heat), and radiative heating. Potential temperature has units of K or $^{\circ}C$, and can be found if you know the air temperature, T , at pressure-level, P , using the following equation

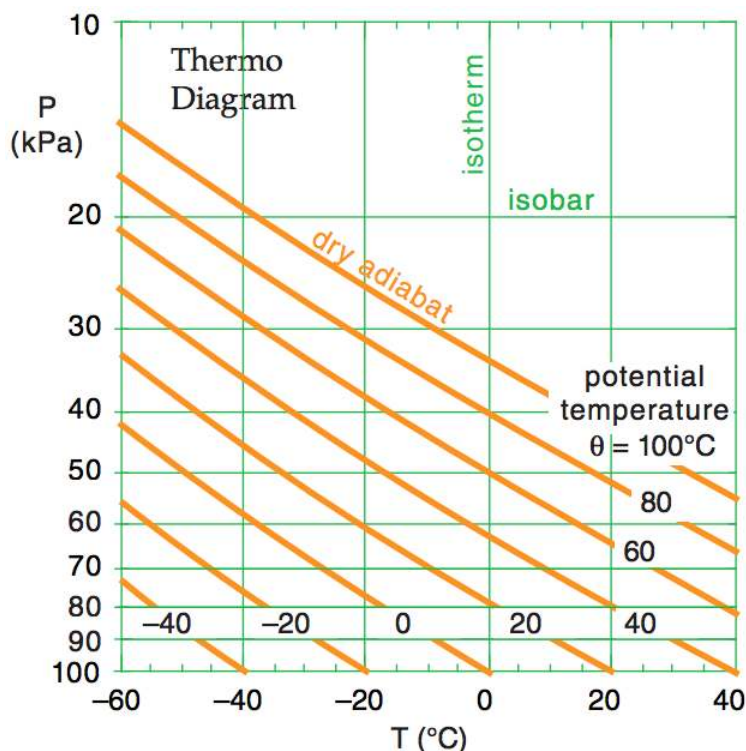
$$\theta = T * \frac{P_0^{R_d/C_p}}{P}$$

where R_d/C_p is a constant equal to 0.28571, and has no units, and P_0 is a reference pressure, typically 1,000 hPa (100,000 Pa), or the local surface pressure.

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3.7: Introduction to Thermodynamic Diagrams

Potential temperature is also very useful in thermodynamic diagrams, which will be briefly introduced here, but covered in more detail in Chapter 5. Thermodynamic diagrams are useful in diagnosing the state of the atmosphere and the buoyancy of air parcels by comparing the temperature difference ΔT between the parcel and its environment. Parcels that are warmer than their environment will tend to rise due to lower density, and the change of the parcel's temperature with height can be anticipated based on the parcel's moisture content. In the thermodynamic diagrams you will use, dry adiabat lines will be plotted to show the dry adiabatic lapse rate, as parcels will cool at this rate until condensation occurs within the parcel. Dry adiabats are labelled with θ because θ is constant along these lines. A simple example is provided here.



Sample thermodynamic diagram ([CC BY-NC-SA 4.0](#)). Horizontal lines show isobars, lines of constant pressure; vertical lines show isotherms, lines of constant temperature; and orange lines show dry adiabats, lines of constant potential temperature.

Heat Budget at Earth's Surface

Before moving on from thermodynamics, let's add another layer of complication to our understanding of Earth's surface heat budget. In Chapter 2 we discussed how the Earth's heat budget could be defined by the incoming shortwave radiation ($K \downarrow$), reflected shortwave radiation ($K \uparrow$), longwave radiation emitted by the Earth ($I \uparrow$), and the downwelling longwave radiation emitted from the atmosphere ($I \downarrow$) received by the Earth's surface.

$$F^* = K \uparrow + K \downarrow + I \uparrow + I \downarrow$$

Earth's surface is considered to be infinitesimally thin with no volume, and no heat can be stored, so the sum of all incoming and outgoing heat fluxes at the surface must balance. The net heat flux at the surface must be zero. In addition to the incoming and outgoing shortwave and longwave radiation, there are three other fluxes that must be considered. But first, let's discuss what is meant by a "flux" of heat.

Heat Flux

Let's say you have a cube of air, somewhere fixed relative to the ground. From an Eulerian framework (fixed-location), pressure changes can be neglected in the First Law equation, as they will be small and slow.

$$\Delta q = C_p \times \Delta T - \frac{\Delta P}{\rho} = C_p \times \Delta T - \frac{0}{\rho} = C_p \times \Delta T$$

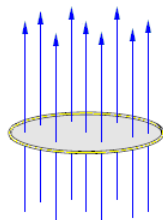
Leftover, you have an equation that states that the heat you transfer to the cube (per unit mass) causes the temperature change:

$$\Delta T = \frac{\Delta q}{C_p}$$

If you divide this by a time interval Δt (note: lower-case t is used for time, T is temperature), gives an equation for temperature change with time:

$$\frac{\Delta T}{\Delta t} = \left(\frac{1}{C_p} \right) \frac{\Delta q}{\Delta t}$$

The temperature of the air cube could be increased if there were a heat transfer into it. Heat flux is the rate of heat transfer through a surface over time, as illustrated in the below figure. The units of heat flux can be given as $\text{J}\cdot\text{m}^{-2}\cdot\text{s}^{-1}$, or $\text{W}\cdot\text{m}^{-2}$, because $\text{W} = \text{J}\cdot\text{s}^{-1}$. It is heat moving through an area over time.



This figure represents a flux of some material or energy through a surface or area ([CC BY-SA 3.0](#)).

Temperature could be increased by heat flux into the cube of air, and could also be decreased by heat flux out of the cube. For example, if there is a net heat flux into a cube of air there will be a net heating effect because heat will be transferred into the cube more quickly than it will leave the cube.

Earth's Surface Budget

Fluxes are defined as positive for heat moving upward. In addition to the net radiation (F^*) from shortwave and longwave radiation, the fluxes include:

F^* = the net radiation between the surface and atmosphere, defined above;

F_H = effective surface turbulent heat flux (sensible heat flux, SH);

F_E = effective surface latent heat flux, caused by evaporation or condensation (latent heat flux, LH); and

F_G = molecular heat conduction to/from deeper below the surface, basically heat being conducted from nearby molecules.

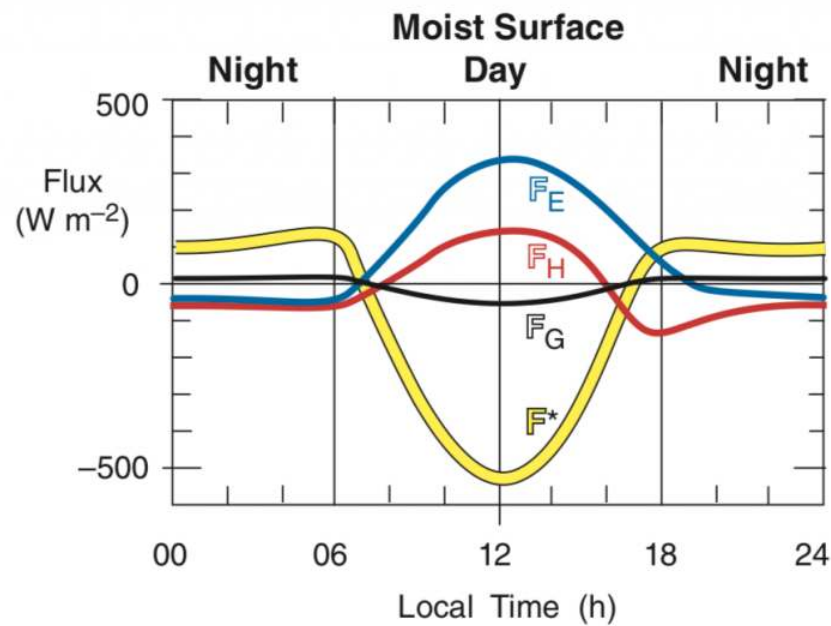
All of these fluxes have to balance.

$$0 = F^* + F_H + F_E - F_G$$

Note

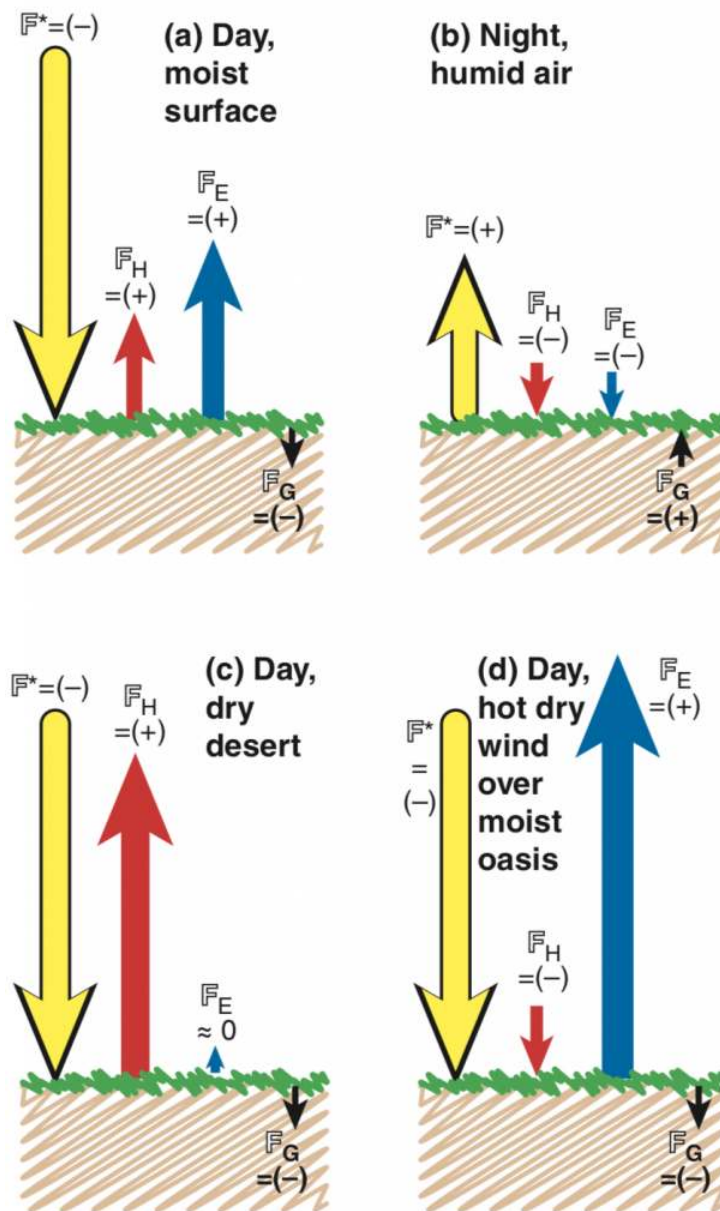
In the class we will likely use SH or SHF to represent the sensible heat flux, and LH or LHF to represent the latent heat flux. Remember, SHF is a dry heat flux from convection and LHF is heat transfer from moisture.

Over the course of a day, the relative contributions from the different terms in the surface heat budget vary. The below figure shows the four terms as they vary over a moist surface during one average day. Notice the yellow line first. There is negative F^* through most of the day as the surface gains heat from the excess shortwave radiation. At night, F^* is positive as the surface radiates infrared radiation upward away from the surface. This is similar to F_G , where heat is transferred downward into the ground during the day and upward at night. However F_H and F_E have opposite signs from F^* and F_G . The surface sensible heat flux, F_E and surface latent heat flux, F_H are positive during the day as convection and evaporation draws heat upward away from the surface.



This figure shows the surface heat balance throughout the day over a moist surface ([CC BY-NC-SA 4.0](https://creativecommons.org/licenses/by-nc-sa/4.0/)).

Now let's look at how the fluxes vary instantaneously over a moist (a, b) or dry (c, d) surface during the day (a, c) and night (b, d). The size of each arrow corresponds to the strength of each type of flux.



This diagram is a representation of Earth's surface fluxes with different sized arrows showing the magnitude of each under different conditions ([CC BY-NC-SA 4.0](#)).

In the above image, note the difference between (a) and (c). Both have a large incoming radiation F^* , but the moist surface (a) has a larger latent heat flux F_E as compared to the dry desert in (c) with a larger sensible heat flux F_H . This shows that heat is transferred from the ground to the atmosphere through evaporation when the surface is moist, but through convection when the surface is dry.

The Bowen ratio helps to distinguish various types of surfaces. The Bowen ratio is defined as

$$BR = \frac{F_H}{F_E} = \frac{SHF}{LHF},$$

the ratio between the sensible heat flux and the latent heat flux. Moist surfaces have a small Bowen ratio because latent heating dominates over sensible heating while dry surfaces have a large Bowen ratio because sensible heating dominates over latent heating.

What other surface heat flux changes do you notice with day or night? How does energy partitioning change depending on the available moisture?

Remember our learning goals for this chapter:

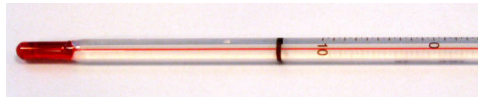
1. Define and describe four methods of energy transfer
2. Describe the change in energy associated with changes in water state
3. Define and apply the first law of thermodynamics
4. Differentiate Eulerian and Lagrangian frameworks
5. Describe the importance of the dry adiabatic lapse rate, and recall what sets its constant value in the atmosphere
6. Compute potential temperature and apply the conserved variable approach
7. Draw a diagram of surface heat fluxes and Earth's radiation budget
8. Compute the Bowen ratio, and define latent and sensible heat flux

Do you feel comfortable with all of the goals?

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3.8: Additional Information

Temperature has been historically measured by a thermometer like the one below. A bulb of expandable fluid grows or shrinks depending on the temperature, indicating the temperature by its top. Thermometers of this type are robust and cheap, but not easily automated.



Liquid in glass thermometer ([Public Domain](#)).

Another type of early thermometer records the minimum and maximum with mercury (below). A marker is pushed upward (for the maximum) or downward (for the minimum).



Minimum and maximum mercury thermometer ([CC BY-SA 3.0](#)).

Today, most thermometers are electrical. They use components that are temperature sensitive, like a temperature sensitive resistor or capacitor where the electrical current changes depending on temperature. Based on the electrical reading from the instrument, temperature can be automatically recorded.



Electrical thermometers are used in systems such as this Elko Automated Surface Observing System (ASOS) ([CC BY-SA 3.0](#)).

This type of thermometer is used in radiosondes, which are launched on weather balloons for measuring temperature throughout the entire atmospheric column. We can observe in-situ as high as 50 km, halfway through the stratosphere with a radiosonde balloon.



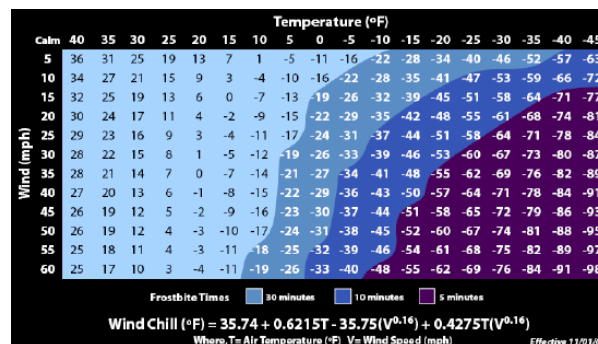
Radiosondes contain thermistor temperature sensors ([CC BY-SA 3.0](#)).

Thermometers need to be correctly sheltered so that solar radiation does not affect the temperature reading. A shelter is typically placed at least 1 meter above the ground and covered in a white housing that blocks the sun but allows air to flow through like the one in the image.



These meteorological shelters protect the instruments inside while still allowing a fairly accurate collection of data ([CC BY 3.0](#)).

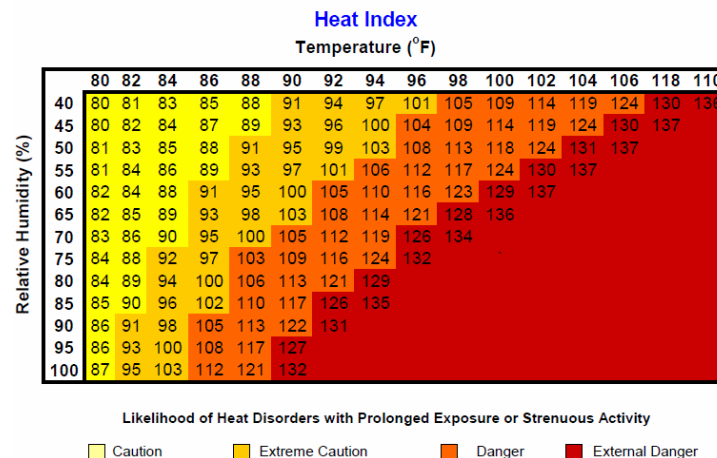
A temperature reading simply gives the temperature of the air, but wind and humidity can affect the way the atmosphere feels to the human body. The below image shows a wind chill graph where the wind speed and the temperature gives an additional measure of temperature called the wind chill. When the air temperature is cold and the wind is high, the temperature feels even colder and can give a person standing outdoors frostbite within minutes (see the purple area below).



Wind chill temperatures for corresponding dry bulb temperatures and wind speed ([Public Domain](#)).

On the other hand when the air is humid, it feels hotter. This is because when the air contains more water vapor and surfaces (like your skin) cannot evaporate it efficiently. Unable to get rid of heat through the latent heat flux (LHF), the surface heats up. This is called the heat index. Notice that a temperature of 84°F at 90% relative humidity has a heat index of 98°F.

NOAA's National Weather Service



The heat index measures how hot it feels compared to the actual air temperature by taking into account the humidity ([Public Domain](#)).

You have experienced at least one of these issues (wind chill or heat index) in your lifetime.

Chapter 3: Questions to Consider

1. Read the following scenarios and chose the type of energy transfer described.
2. Fill in the blanks to describe what happens when an air parcel rises.
2. Earlier in the chapter, a brick being held up was described as having potential energy. If the brick weighs 3.5 kg and is being held 1.2 m off the ground, how much potential energy does it have? Remember that acceleration from gravity is $9.8 \text{ m}\cdot\text{s}^{-2}$.
3. Mauna Kea, an inactive volcano on the island of Hawai'i, has an elevation of 13,803 ft at the summit. If a weather balloon measures the environmental lapse rate to be $6.5 \text{ K}\cdot\text{km}^{-1}$, how much cooler is the summit than the rest of the island at sea level?
4. In the winter, it often snows on top of Mauna Kea. Explain how that's possible despite the mountain's tropical location using your result from the previous question.

Selected Practice Question Answers:

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CHAPTER OVERVIEW

4: Water Vapor

Alison Nugent and Shintaro Russell

Learning Objectives

By the end of this chapter, you should be able to:

- Compute saturation vapor pressure using the Clausius-Clapeyron equation;
- Convert between humidity variables. Differentiate between relative humidity, specific humidity, absolute humidity, wet-bulb temperature, mixing ratio, and dew point;
- Describe the conditions for saturation to occur;
- Apply the moist adiabatic lapse rate;
- Use the principles of phase change and latent heating to describe why the moist adiabatic lapse rate is less than the dry adiabatic lapse rate.

[4.1: Introduction](#)

[4.2: Saturation](#)

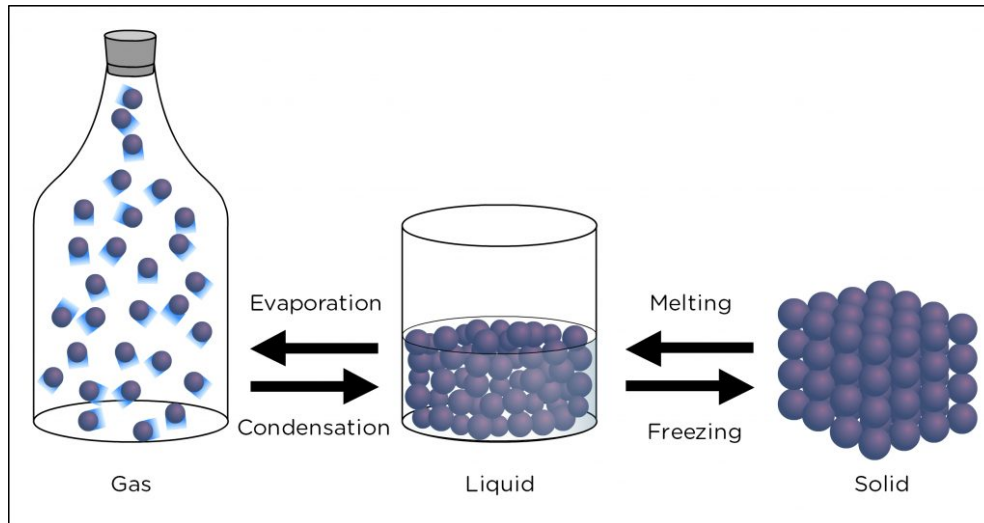
[4.3: Humidity Variables](#)

[4.4: Why Do We Care So Much About Moisture?](#)

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4.1: Introduction

Water can exist as a solid, liquid, or gas at typical conditions found on Earth. As we learned, the process of liquid water becoming water vapor is called evaporation and this process absorbs or requires energy. The opposite process is called condensation, where water vapor becomes liquid water, releasing energy. Condensation is especially important in atmospheric science because this is the process that allows clouds to form.



Phase changes of water from gas (water vapor) to liquid (water) to solid (ice) with the names for the processes also labeled. ([CC BY 2.0](#)).

Clouds are composed of millions and billions of tiny liquid water droplets. How do they form? Why are they there?



Water droplets condensed on a glass surface ([CC BY 2.0](#)).

Before we can understand clouds in the atmosphere, we need to explore concepts like how humidity is defined and what saturation means.

In general, humidity is the amount of water vapor in the air. You've likely heard of relative humidity and dew point temperature, but what do these quantities mean physically?

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4.2: Saturation

Imagine a closed jar filled halfway with water. At the initial time, more water molecules evaporate from the water surface than the number that return. However, after some time, the number of molecules evaporating from the surface will be equal to the number of molecules condensing back into the water surface. When condensation and evaporation are equal, this is called **saturation**.

Saturation occurs when air contains the maximum amount of water vapor possible for its given temperature. That is why condensation equals evaporation. If evaporation occurs, the air cannot contain more water vapor, so some must condense. Now let's get quantitative.

Vapor pressure at saturation

Every gas in the atmosphere exerts pressure, for example, vapor pressure makes up a fraction of the total atmospheric pressure. In the following equation, all of the gases in Earth's atmosphere contribute to the total atmospheric pressure $P_{\text{atmosphere}}$.

$$P_{\text{atmosphere}} = P_{N_2} + P_{O_2} + P_{Ar} + \dots P_{H_2O}$$

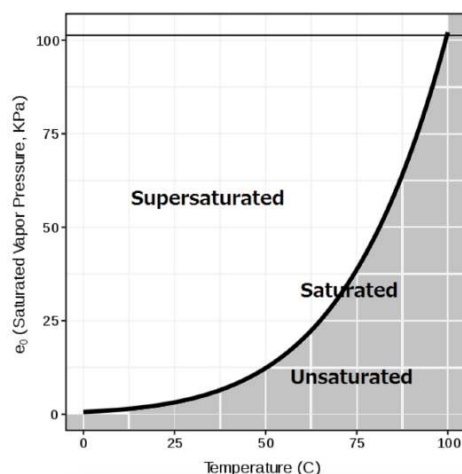
Specifically for water vapor, the more water vapor that is added to the atmosphere, the higher the vapor pressure P_{H_2O} . The units for vapor pressure are the same as pressure and can be in Pascals, hectoPascals, or kiloPascals. Because we are staying consistent with Roland Stull's *Practical Meteorology* textbook, we will use kiloPascals (kPa) throughout this chapter.

The amount of water vapor that the atmosphere can contain depends on temperature. Lower temperature air cannot contain as much water vapor as higher temperature air. If we think of this quantitatively in terms of pressure, *saturation vapor pressure* refers to the pressure exerted by the movement of water vapor molecules exerted over a surface of liquid water. When the partial pressure exerted by water vapor is equal to the saturation vapor pressure, the air is saturated.

The **Clausius-Clapeyron equation** gives the approximate relationship between saturation vapor pressure (e_s) and temperature in the atmosphere

$$e_s \approx e_0 \cdot \exp \left[\frac{L_v}{R_v} \cdot \left(\frac{1}{T_0} - \frac{1}{T} \right) \right]$$

where the water-vapor gas constant R_v is $461 \text{ J}\cdot\text{K}^{-1}\cdot\text{kg}^{-1}$, T_0 is 273.15 K, e_0 is 0.6113 kPa, and L_v is the latent heat of vaporization, $2.5 \times 10^6 \text{ J}\cdot\text{kg}^{-1}$. This results in L_v/R_v being equal to 5423 K. In this equation, units for temperature must be in Kelvin. Note that in the equation above, $\exp[x]$ implies the exponential function e^x , but it is written on one line for visual purposes.



A graph of the saturation vapor pressure as a function of temperature showing the exponential relationship between the two from the Clausius-Clapeyron equation (Modified from [CC BY-SA 4.0](#)).

The image shows the relationship between temperature and saturation vapor pressure based on the Clausius-Clapeyron equation. Lower temperatures are saturated with respect to water vapor at lower vapor pressures, while higher temperatures need higher vapor pressures to be saturated. Temperature is the primary factor determining water vapor saturation.

In the graph of saturation vapor pressure vs. temperature notice the saturation vapor pressure value at the boiling temperature, 100 °C. The saturation vapor pressure value $e_s(100\text{ °C})=101.325\text{ kPa}$, is the same value as the atmospheric surface pressure. Water boils at the Earth's surface when the saturation vapor pressure is equal to the atmospheric pressure, which is why water boils at 100 °C. Will water boil at the same temperature at the top of Mount Everest?

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4.3: Humidity Variables

Vapor pressure is one way of defining humidity, but there are many others. Here is a non-comprehensive list of humidity variables and their typical units.

e = vapor pressure (kPa)

r = mixing ratio ($\text{g}\cdot\text{kg}^{-1}$)

q = specific humidity ($\text{g}\cdot\text{kg}^{-1}$)

ρ_V = absolute humidity ($\text{g}\cdot\text{m}^{-3}$)

RH = relative humidity (%)

z_{LCL} = lifting condensation level (km)

T_d = dew point (temperature) ($^{\circ}\text{C}$)

T_w = wet-bulb temperature ($^{\circ}\text{C}$)

Vapor Pressure

We've already discussed saturation vapor pressure, e_s , but you can also compute vapor pressure, e . However, because T_d is often unknown, the easiest way is usually through relative humidity.

$$e = e_0 \cdot \exp \left[\frac{L_v}{R_v} \cdot \left(\frac{1}{T_0} - \frac{1}{T_d} \right) \right]$$

Again, e_0 is 0.6113 kPa, L_v is $2.5 \times 10^6 \text{ J}\cdot\text{kg}^{-1}$, R_v is $461 \text{ J}\cdot\text{K}^{-1}\cdot\text{kg}^{-1}$, T_0 is 273.15 K, and T_d is dew point temperature, which will be defined later.

Mixing Ratio

Mixing ratio, r , is the ratio of the mass of water vapor to the mass of dry air. It is typically expressed as grams of water vapor per kilogram of air ($\text{g}\cdot\text{kg}^{-1}$).

$$r = \frac{m_{\text{water vapor}}}{m_{\text{dry air}}}$$

$$r = \frac{\epsilon \cdot e}{P - e}$$

Pressure (P) should be in the same units as vapor pressure (e). The constant ϵ is 0.622, is the ratio between the gas constant for dry air and the gas constant for water vapor.

$$\epsilon = \frac{R_d}{R_v} = \frac{287 \text{ J kg}^{-1} \text{ K}^{-1}}{461 \text{ J kg}^{-1} \text{ K}^{-1}} = 0.622$$

Saturation mixing ratio, r_s , is computed the same way as the mixing ratio but with saturation vapor pressure, e_s , instead of e .

$$r_s = \frac{\epsilon \cdot e_s}{P - e_s}$$

When calculating mixing ratio, the pressure units on the top of the fraction will cancel with the pressure units on the bottom of the fraction. While it appears unit-less, it's technically not based on its definition of mass of water vapor as compared to mass of dry air. See the Pro Tip below for more information.

Pro Tip: Many units of moisture are given in $\text{g}\cdot\text{kg}^{-1}$ or $\text{kg}\cdot\text{kg}^{-1}$ so technically the units could cancel and it could be unitless! Don't let this fool you. It is important to remember that the mass in the numerator and denominator are different. In the case of mixing ratio, the value is given in mass of water vapor proportional to the mass of dry air.

Specific Humidity

Specific humidity, q , is the ratio of the mass of water vapor to the total mass of air (dry air and water vapor combined). It is expressed as grams of water vapor per kilogram of air ($\text{g}\cdot\text{kg}^{-1}$).

$$q = \frac{m_{\text{water vapor}}}{m_{\text{total air}}}$$

$$q = \frac{\epsilon \cdot e}{P - e \cdot (1 - \epsilon)} \approx \frac{\epsilon \cdot e}{P}$$

Again, saturation specific humidity, q_s , is computed with e_s , instead of e .

$$q_s = \frac{\epsilon \cdot e_s}{P - e_s \cdot (1 - \epsilon)} \approx \frac{\epsilon \cdot e_s}{P}$$

Absolute Humidity

Absolute humidity, ρ_v , is the ratio of the mass of water vapor to the volume of air. It is expressed as grams of water vapor in a cubic meter of air ($\text{g} \cdot \text{m}^{-3}$). It is effectively water vapor density.

$$\rho_v = \frac{m_{\text{water vapor}}}{\text{Volume}}$$

$$\rho_v = \frac{e}{R_v \times T}$$

Again, saturation absolute humidity, ρ_{vs} , uses e_s , instead of e .

$$\rho_{vs} = \frac{e_s}{R_v \times T}$$

Relative Humidity

Relative humidity, RH , is the ratio of the amount of water vapor present in the air to the maximum amount of water vapor needed for saturation at a certain pressure and temperature. It is typically multiplied by 100 and expressed as a percent. Relative humidity shows how close the air is to being saturated, not how much water vapor the air contains. For this reason, RH is not a good indicator of the quantitative amount of water vapor in the air. It is only a relative measure that is highly dependent on the air temperature. Relative humidity greater than 100% is called supersaturation.

$$RH = \frac{e}{e_s} = \frac{RH\%}{100\%}$$

or

$$RH = \frac{q}{q_s} = \frac{\rho_v}{\rho_{vs}}$$

Imagine two parcels of air with the same volume, pressure, and relative humidity. Parcel 1 has an air temperature of 20 °C while Parcel 2 has an air temperature of 30 °C. Which parcel contains more water vapor?

Dew Point Temperature

The dew point temperature, T_d , is the temperature to which the air must be cooled to reach saturation, without changing the moisture or air pressure. It measures the actual moisture content of a parcel of air. Saturation occurs when the dew point temperature equals the air temperature.

$$T_d = \left[\frac{1}{T_0} - \frac{R_v}{L_v} \cdot \ln \left(\frac{e}{e_0} \right) \right]^{-1}$$

$$T_d = \left[\frac{1}{T_0} - \frac{R_v}{L_v} \cdot \ln \left(\frac{r \cdot P}{e_0 \cdot (r + \epsilon)} \right) \right]^{-1}$$

When the dew point temperature is lower than the freezing point of water, it is also called the *frost point*.

Wet-Bulb Temperature

Wet-bulb temperature, T_w , is the lowest temperature that can be achieved if water evaporates within the air. When the relative humidity is 100%, the wet-bulb temperature is equal to the air temperature because there is no evaporation.

The wet-bulb temperature is difficult to calculate but easy to measure. To measure the wet-bulb temperature, all you need is a thermometer with a wet cloth wrapped around the bulb. Typically this thermometer is attached to an apparatus called a sling psychrometer to make it easy to spin around in the air to create lots of airflow over the wet cloth on the thermometer. The evaporation from the wet cloth cools the temperature measured, hence the wet-bulb temperature is always lower than the air temperature (or dry-bulb temperature) when relative humidity is less than 100%.

You can also estimate the wet bulb temperature using lines on a graph. *Normand's Rule* is used to calculate the wet-bulb temperature from the air temperature and the dew point temperature. The wet-bulb temperature is always between the dew point and the dry-bulb temperature ($T_d \leq T_w \leq T$). This can be implemented on thermodynamic diagrams, such as the Skew- $T \log P$, which is discussed in more detail in the next chapter.

Take note of this description for later. To find the wet-bulb temperature on a Skew- $T \log P$ diagram, follow the dry adiabatic lapse rate line upward from the air temperature. Next, use the dew point temperature and follow an *isohume* (line of constant relative humidity) upward. The point where these two lines meet is called the lifting condensation level (LCL). From the meeting point, follow the moist (saturated) adiabatic lapse rate back down to obtain the wet-bulb temperature value. This is probably confusing at this point because we have not discussed the LCL or the moist adiabatic lapse rate, but don't worry, we'll repeat this logic again in the next chapter to make sure this is clear.

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4.4: Why Do We Care So Much About Moisture?

You may be wondering at this point why we care so much about moisture and why we need so many definitions of (almost) the same thing. The reason is that moisture is an extremely important atmospheric property. Water can exist in three phases (vapor, liquid, ice) within the atmosphere at typical pressures and temperatures. It has an especially large impact on the human experience—think about a humid day, foggy conditions, rain, snow, or even hail! Less obvious is its impact on atmospheric stability, which drives the aforementioned conditions.

For now, let's think about the process of water vapor condensing to form liquid water. There is one final definition of humidity that will be helpful.

Lifting Condensation Level

The lifting condensation level, z_{LCL} , is the altitude where clouds form. At the LCL, temperature equals the dew point temperature, resulting in saturation and therefore condensation. The height (z) of the LCL is

$$z_{LCL} = a \cdot (T - T_d)$$

where a is $0.125 \text{ km } ^\circ\text{C}^{-1}$. We can also define the temperature at the LCL as follows.

$$T_{LCL} = T - \Gamma_d \cdot z_{LCL}$$

Moist Adiabatic Lapse Rate

In the last chapter, we discussed how temperature changes as a dry parcel of air is lifted in the atmosphere. You will recall that as an air parcel is lifted, the temperature drops by 9.8 K every km due to the work the air parcel must do to the environment as it expands. Let's add moisture to the discussion and see how this changes things.

If the air parcel reaches saturation (100% relative humidity) and water vapor condenses to liquid water within the parcel, latent heat will be released. In the case of a rising air parcel that is cooling from adiabatic expansion, this added heat from condensation counterbalances some of the cooling. Hence, the air parcel will no longer cool at the dry adiabatic lapse rate but at the smaller moist adiabatic lapse rate (Γ_m). Unlike the dry adiabatic lapse, the moist adiabatic lapse rate is not constant and varies based on the temperature and moisture of the air parcel.

We will approximate the moist adiabatic lapse rate with the following value.

$$\Gamma_m = 4.5 \frac{K}{km} = 4.5 \frac{^\circ\text{C}}{km}$$

The difference between the dry adiabatic lapse rate (Γ_d) and the moist adiabatic lapse rate (Γ_m) is significant and has profound influence on atmospheric stability, the topic of the following chapter.

Chapter 4: Questions to Consider

1. Explain the conditions needed for saturation to occur.
2. What is the saturation vapor pressure of air at 26°C?
3. Explain the difference between specific humidity and relative humidity.
4. If the temperature is 10°C and the pressure is 700 hPa, calculate the saturation specific humidity and the saturation mixing ratio.
5. Explain why the moist adiabatic lapse rate is less than the dry adiabatic lapse rate.

Selected Practice Question Answers:

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CHAPTER OVERVIEW

5: Atmospheric Stability

Alison Nugent and David DeCou

Learning Objectives

By the end of this chapter, you should be able to:

- Interpret stability based on the dry and moist adiabatic lapse rates
- Understand how stability relates to vertical motion in the atmosphere
- Describe and differentiate between the many lines on a Skew-T log-P diagram
- Find the LCL, regions of CAPE and CIN, and the tropopause from a Skew-T log-P diagram

[5.1: Introduction](#)

[5.2: Atmospheric Stability and Lapse Rates](#)

[5.3: Skew-T Log-P Diagram](#)

[5.4: CAPE and CIN](#)

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5.1: Introduction



A cumulonimbus cloud rises and expands as a product of instability within the atmosphere ([Public Domain](#)).

When you think of the word “stable,” you typically think of an object that is unlikely to change or something that is balanced. The opposite is true with something that is “unstable”. An unstable object is likely to fall or change position with time. The same is true with clouds. When you see a fluffy cumulus cloud, you might notice them changing shape from one minute to the next. Such clouds are in a constant state of change, and thus represent the atmosphere in an unstable state.

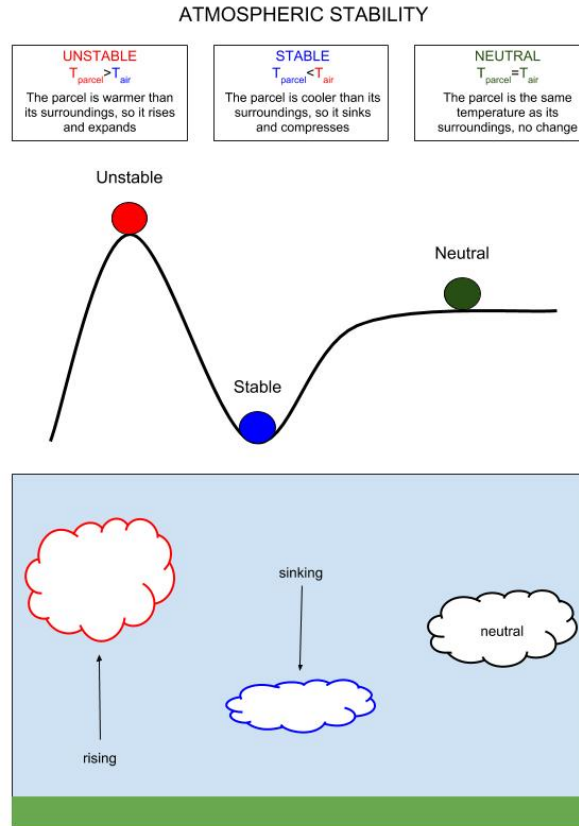


A perfect cumulus cloud just west of Magic Island (photo by Sarah Williamson).

Instability in the atmosphere is a concept that is intimately connected with thunderstorms, cumulus development, and vertical motion. In order to visualize the concept of **stability**, you might imagine a boulder sitting at the bottom of a canyon surrounded by steep hills, as depicted in the figure below by the blue circle. If you were strong enough to push the boulder from its initial position partway up one of the hills, it would roll back to the bottom once you let go. Despite exerting a force on the boulder and causing an initial displacement, it would return to its initial position, and the net displacement would be zero. To visualize the concept of **instability**, imagine the same boulder at the top of a hill (red circle below). If you were able to push the boulder just a little bit in any direction, it would begin to roll downward and accelerate away from its initial position. However, if the same boulder were to be placed on flat ground (green circle below) and you were to push it, it would change position, but remain in its new position. This is an example of **neutral stability**.

Each of these concepts can be applied to motions of air parcels in the atmosphere. The topic of stability in atmospheric science is important because the formation of clouds is closely related to stability or instability in the atmosphere. In this chapter we will

connect these concepts to the buoyancy of air parcels, and learn to use thermodynamic diagrams to visualize movement.



Examples of stability and instability in relation to air and parcel temperatures (created by Britt Seifert).

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5.2: Atmospheric Stability and Lapse Rates

Adiabatic Processes

When discussing stability in atmospheric sciences, we typically think about air parcels, or imaginary blobs of air that can expand and contract freely, but do not mix with the air around them or break apart. The key piece of information is that movement of air parcels in the atmosphere can be estimated as an adiabatic process. **Adiabatic processes** do not exchange heat and they are reversible.

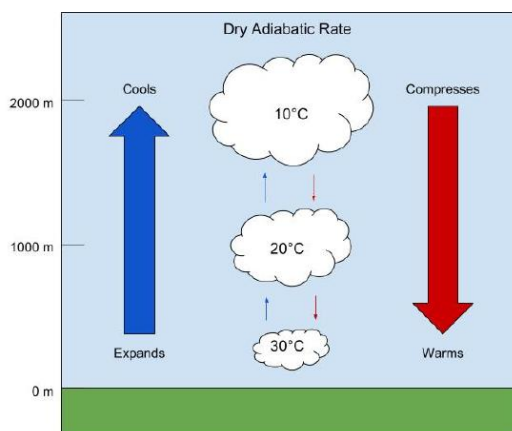
Imagine you have a parcel of air at the Earth's surface. The air parcel has the same temperature and pressure as the surrounding air, which we will call the environment. If you were to lift the air parcel, it would find itself in a place where the surrounding environmental air pressure is lower, because we know that pressure decreases with height. Because the environmental air pressure outside the parcel is lower than the pressure inside the parcel, the air molecules inside the parcel will effectively push outward on the walls of the parcel and expand adiabatically. The air molecules inside the parcel must use some of their own energy in order to expand the air parcel's walls, so the temperature inside the parcel decreases as the internal energy decreases. To summarize, rising air parcels expand and cool adiabatically without exchanging heat with the environment.

Now imagine that you move the same air parcel back to Earth's surface. The air parcel is moving into an environment with higher air pressure. The higher environmental pressure will push inward on the parcel walls, causing them to compress, and raise the inside temperature.

The process is adiabatic, so again, no heat is exchanged with the environment. However, temperature changes in the air parcel can still occur, but it is not due to mixing, it is due to changes in the internal energy of the air parcel.

Dry Adiabatic Lapse Rate

As long as an air parcel is unsaturated (relative humidity < 100%), the rate at which its temperature will change will be constant. A decrease in temperature with height is called a **lapse rate** and while the temperature decreases with altitude, it is defined as positive because it is a *lapse* rate. Recall from chapter 3 that the **dry adiabatic lapse rate**, Γ_d , is equal to $9.8 \text{ K} \cdot \text{km}^{-1} = 9.8 \text{ }^{\circ}\text{C} \cdot \text{km}^{-1}$. This drop in temperature is due to adiabatic expansion and a decrease in internal energy.



Air rises, expands, and cools at the dry adiabatic lapse rate, approximated as a 10°C decrease per km (created by Britt Seifert).

Let's get back to the topic of atmospheric stability. Stability in the atmosphere refers to a condition of equilibrium. As discussed with the example of the boulder on a hill or valley, some initial movement resulted in either more (unstable), less (stable), or no change (neutral). Given some initial change in the elevation of an air parcel, if the air is in stable equilibrium, the parcel will tend to return back to its original position after it is forced to rise or sink. In an unstable equilibrium, an air parcel will accelerate away from its initial position after being pushed. The motion could be upward or downward, but generally unstable atmospheres favors vertical motions. Finally, in a neutral equilibrium, some initial change in the elevation of an air parcel will not result in any additional movement.

Determining Stability

How do you know if an air parcel will be stable after some initial displacement? Stability is determined by comparing the temperature of a rising or sinking air parcel to the environmental air temperature. Imagine the following: at some initial time, an air parcel has the same temperature and pressure as its environment. If you lift the air parcel some distance, its temperature drops by $9.8 \text{ K}\cdot\text{km}^{-1}$, which is the dry adiabatic lapse rate. If the air parcel is colder than the environment in its new position, it will have higher density and tend to sink back to its original position. In this case, the air is stable because vertical motion is resisted. If the rising air is warmer and less dense than the surrounding air, it will continue to rise until it reaches some new equilibrium where its temperature matches the environmental temperature. In this case, because an initial change is amplified, the air parcel is unstable. In order to figure out if the air parcel is unstable or not we must know the temperature of both the rising air and the environment at different altitudes.

One way this is done in practice is with a weather balloon. We can get a vertical profile of the environmental lapse rate by releasing a radiosonde attached to a weather balloon. A radiosonde sends back data on temperature, humidity, wind, and position, which are plotted on a thermodynamic diagram. This vertical plot of temperature and other variables is known as a **sounding**.

Dry Stability

If an air parcel is dry, meaning unsaturated, stability is relatively straightforward. An atmosphere where the environmental lapse rate is the same as the dry adiabatic lapse rate, meaning that the temperature in the environment also drops by $9.8 \text{ K}\cdot\text{km}^{-1}$, will be considered neutrally stable. After some initial vertical displacement, the temperature of the air parcel will always be the same as the environment so no further change in position is expected.

If the environmental lapse rate is less than the dry adiabatic lapse rate, some initial vertical displacement of the air parcel will result in the air parcel either being colder than the environment (if lifted), or warmer than the environment (if pushed downward). This is because if lifted, the temperature of the air parcel would drop more than the temperature of the environment. This is a stable situation for a dry air parcel and a typical scenario in the atmosphere. The **global average tropospheric lapse rate is $6.5 \text{ K}\cdot\text{km}^{-1}$** , which is stable for dry lifting.

Finally, if the environmental lapse rate is greater than the dry adiabatic lapse rate, some initial vertical displacement of the air parcel will result in the air parcel either being warmer than the environment (if lifted), or colder than the environment (if pushed downward). This is because if lifted, the temperature of the air parcel would drop less than the temperature of the environment. This is an unstable situation for a dry air parcel.

In general for a dry air parcel, the following is true.

$$\Gamma_d = \Gamma_{env} \quad \text{NEUTRAL}$$

$$\Gamma_d < \Gamma_{env} \quad \text{STABLE}$$

$$\Gamma_d > \Gamma_{env} \quad \text{UNSTABLE}$$

Moist Adiabatic Lapse Rate

When moisture is added, everything gets more complicated. In Chapter 4 we learned that whether or not an air parcel is saturated depends primarily on its temperature and, of course, its moisture content. The graph of the Clausius-Clapeyron relationship shows us that given the same amount of moisture, air is more likely to be saturated at a lower temperature.

We know that as an air parcel is lifted, its temperature drops according to the dry adiabatic lapse rate. So what happens when the air parcel is cold enough that the air becomes saturated with respect to water vapor? The short answer is that if it continues to cool, water vapor will condense to liquid water to form a cloud.

When water vapor condenses, it goes from a higher energy state to a lower energy state. Energy is never created nor destroyed, especially in phase changes, so what happens to all that excess energy? The energy gets released in the form of latent heat. The latent heat of condensation is approximately equal to $2.5 \times 10^6 \text{ J}\cdot\text{kg}^{-1}$, which means that for every kg of water vapor that condenses to form liquid water, 2.5×10^6 Joules of energy are released.

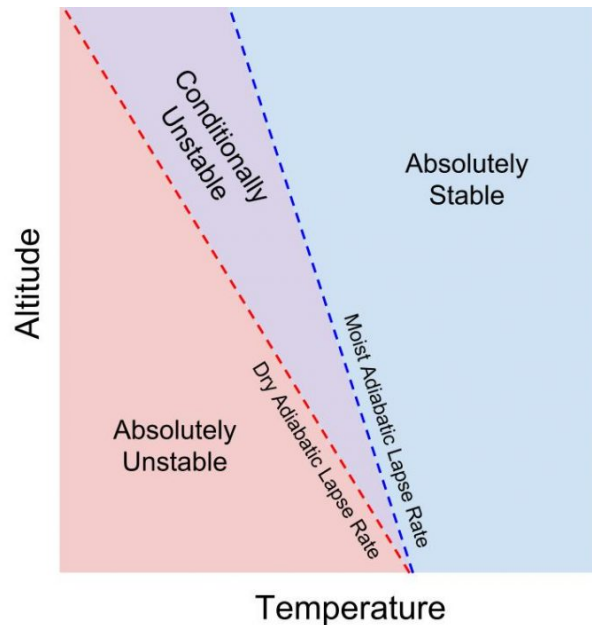
This has large consequences for the lapse rate of an air parcel and distinguishes the dry adiabatic lapse rate from the moist adiabatic lapse rate. As latent heat is added from the process of condensation, it offsets some of the adiabatic cooling from expansion.

Because of this, the air parcel will no longer cool at the dry adiabatic lapse rate, but will cool as a slower rate, known as the **moist adiabatic lapse rate**. To summarize, a parcel will cool at the dry adiabatic rate until it is saturated, after which it won't cool as quickly due to condensation. The moist adiabatic lapse rate varies a little by temperature, but in this class we will consider it a constant for simplicity:

$$\Gamma_m = 4.5 \text{ K} \cdot \text{km}^{-1} = 4.5^\circ \text{C} \cdot \text{km}^{-1}$$

Moist Stability

The effects of moisture change the lapse rate of the air parcel and, therefore, affects stability. However, the concepts are still the same and we still compare the air parcel temperature to the environmental temperature. We have just one added complication to worry about—we need to know whether the air parcel is dry or moist. Some definitions are included below, which take into account both dry and moist adiabatic lapse rates.



A thermodynamic diagram showing the stability of the atmosphere based on the dry ($\Gamma_d = 9.8 \text{ K km}^{-1}$) and moist ($\Gamma_m = 4.5 \text{ K km}^{-1}$) adiabatic lapse rates (Created by Britt Seifert).

The atmosphere is said to be **absolutely stable** if the environmental lapse rate is less than the moist adiabatic lapse rate. This means that a rising air parcel will always cool at a faster rate than the environment, even after it reaches saturation. If an air parcel is cooler at all levels, then it will not be able to rise, even after it becomes saturated (when latent heating will counteract some cooling).

The atmosphere is said to be **absolutely unstable** if the environmental lapse rate is greater than the dry adiabatic lapse rate. This means that a rising air parcel will always cool at a slower rate than the environment, even when it is unsaturated. This means that it will be warmer (and less dense) than the environment, and allowed to rise.

The atmosphere is said to be **conditionally unstable** if the environmental lapse rate is between the moist and dry adiabatic lapse rates. This means that the buoyancy (the ability of an air parcel to rise) of an air parcel depends on whether or not it is saturated. In a conditionally unstable atmosphere, an air parcel will resist vertical motion when it is unsaturated, because it will cool faster than the environment at the dry adiabatic lapse rate. If it is forced to rise and is able to become saturated, however, it will cool at the moist adiabatic lapse rate. In this case, it will cool slower than the environment, become warmer than the environment, and will rise.

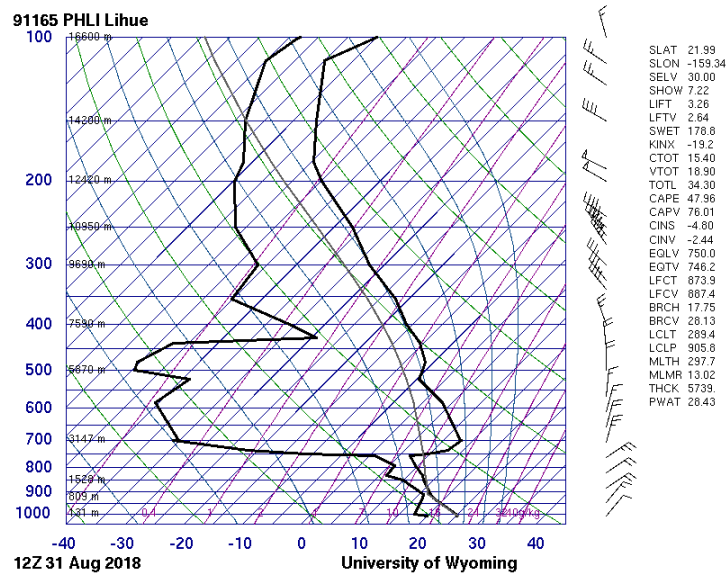
Hawaiian Focus Box

Around Hawaii, the atmosphere is almost always conditionally unstable, meaning that the environmental lapse rate lies somewhere between the dry and moist adiabatic lapse rates. For this reason, Hawaii almost always has convective clouds. Convective clouds are clouds where the edges are bumpy and cumuliform, like cauliflower. The clouds are convective because the atmosphere is stable to dry lifting and unstable to moist lifting. Once the air is saturated, instability sets in and vertical motion takes off. This is especially common as air is lifted over our mountainous islands. The forced lifting from the terrain creates clouds and rain right over the mountains! In scientific terms, the initial lifting of the stable low level dry air by the terrain causes the air to adiabatically expand and reach saturation, at which point the environment is unstable to moist lifting and convection is the result.

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5.3: Skew-T Log-P Diagram

There are many different types of thermodynamic diagrams, but the main one we will discuss are Skew-T Log-P diagrams, so-named because the isotherms (lines of equal temperature, T) on the diagram are slanted (skewed) and the isobars (lines of equal pressure, P) on the diagram are in log space. Here we will focus on how to read and utilize Skew-T Log-P diagrams (often shortened to Skew-T diagram) to determine parcel buoyancy and atmospheric stability.

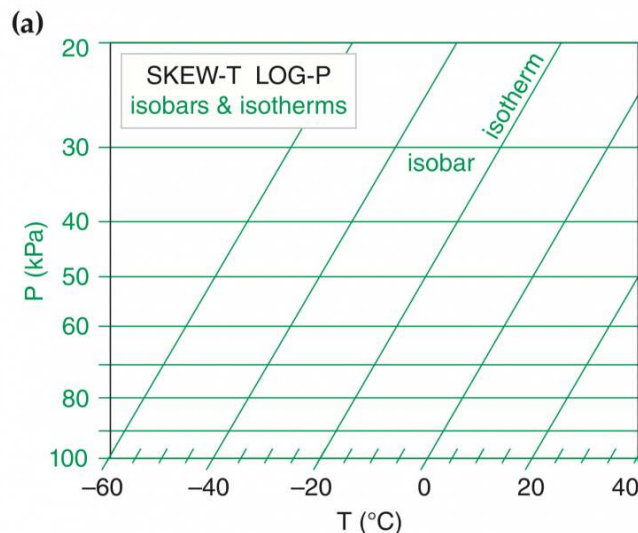


An example Skew-T Log-P diagram from Lihue on August 31st, 2018. The sounding was retrieved from the Upper Air Soundings portion of the University of Wyoming Weather Web: <http://weather.uwyo.edu/upperair/sounding.html> (Copyright 2018 by University of Wyoming Department of Atmospheric Science, used with permission.)

The radiosonde balloon sounding plotted here was launched from Lihue on Kauai (see the top left, labelled as station “91165 PHLI Lihue”). You can see the vertical environmental temperature profile (T) plotted as the black jagged line on the right. The dew point temperature (T_d) with height is plotted with the black jagged line on the left. Although this figure may be overwhelming to read at first, we’ll walk through it together. The horizontal axis is temperature in $^{\circ}\text{C}$, with temperatures increasing to the right. The vertical axis is air pressure in hPa , decreasing with height, so higher heights are toward the top of the chart. When the T and T_d lines are close together, the environment has a high relative humidity and the air is closer to saturation. In this particular sounding, there is a lot of moisture near the surface, but dries out in the mid-levels.

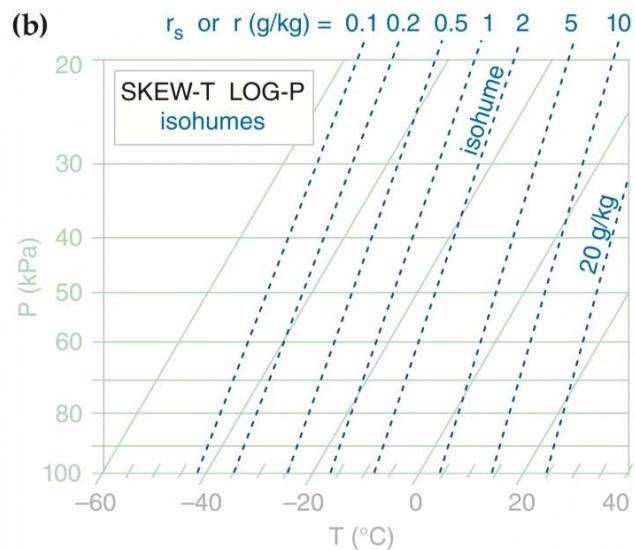
Radiosonde balloons are launched twice a day (00Z and 12Z) from many locations around the world. The latitude and longitude for the station is given in the top of the list on the right where station latitude (SLAT) is given as 21.99 degrees North and SLON is -159.34 degrees West. The station elevation SELV is 30 m. The sounding time and date is given in the bottom left, and the bottom right says “University of Wyoming” because in this particular example, the University of Wyoming is the organization that gathered and archived the dataset. You can find soundings for other locations and dates at this website: <http://weather.uwyo.edu/upperair/sounding.html>.

Let’s go through the lines one by one.



Isobars (horizontal, lines of constant pressure) and isotherms (slanted, lines of constant temperature) ([CC BY-NC-SA 4.0](#)).

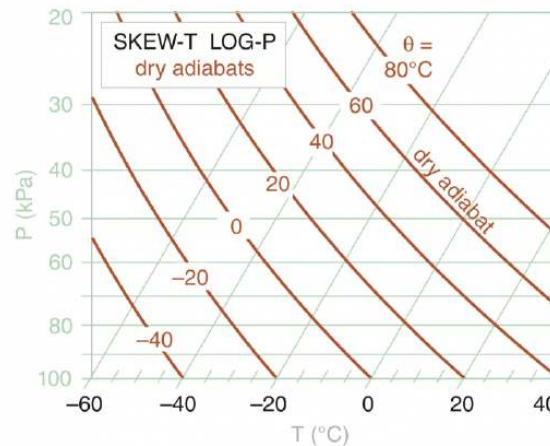
The horizontal lines on a Skew-T are isobars, or lines of equal air pressure. You will typically see them given in hPa , but the lines in the above figure are in kPa . The isobars have larger spaces as you get toward the top of the diagram because they are logarithmic with height. The evenly-spaced solid lines that slant up and to the right are isotherms, or lines of equal temperature (T). This allows colder temperatures to be plotted on the diagram.



Isohumes (slanted dashed lines), lines of constant mixing ratio ([CC BY-NC-SA 4.0](#)).

The dashed lines that run up and to the right are isohumes, or lines of constant mixing ratio. These are typically given in units of $g \cdot kg^{-1}$. If you use a Skew-T where these lines are not dashed or color-coded, remember that these are spaced more closely together than isotherms and are more steep. They also do not line up with the temperature labels on the x-axis.

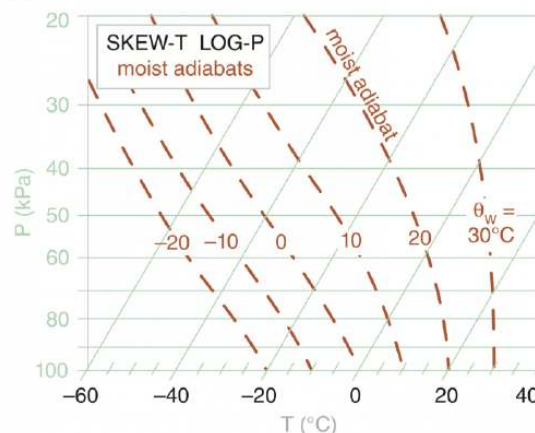
(c)



Dry adiabatic lapse rate reference lines, also known as lines of constant potential temperature ([CC BY-NC-SA 4.0](#)).

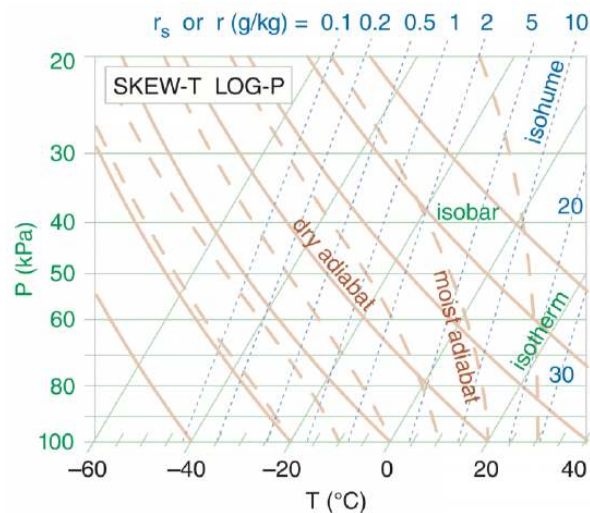
The evenly-spaced curved solid lines that run from bottom right to top left are dry adiabats, and depict the dry adiabatic lapse rate ($9.8 \text{ K}\cdot\text{km}^{-1}$). The dry adiabatic lapse rate is considered a constant, but you can see here that over large changes in temperature and pressure, it varies a little. Don't worry about these variations—we still consider it a constant. Dry adiabatic lapse rate reference lines are also called lines of constant potential temperature (θ). The dry adiabats always curve upward from right to left in a concave way.

(d)



Moist adiabatic lapse rate reference lines. ([CC BY-NC-SA 4.0](#)).

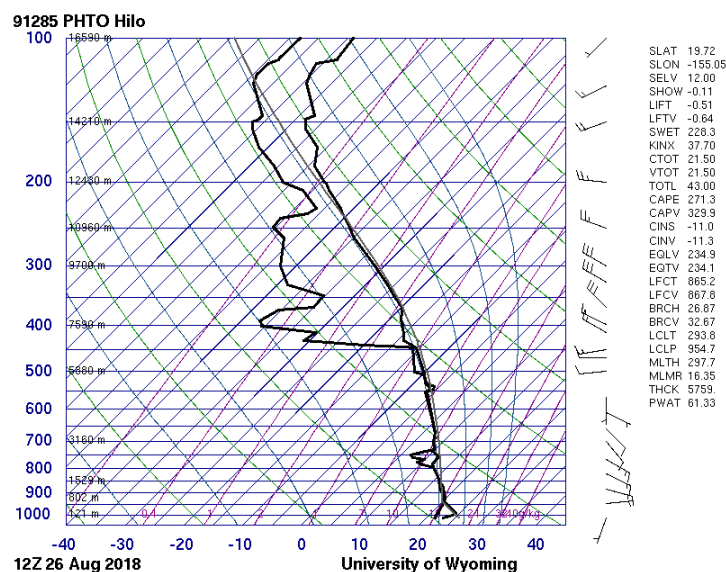
The uneven, dashed, lines that curve up and to the left are the moist adiabats. The moist adiabatic lapse rate varies with both temperature and moisture content, but is close to the dry adiabatic lapse rate at high altitudes due to cold temperatures and small moisture content. These lines are parallel to the dry adiabats higher up on the Skew-T Log-P diagram. These are also lines of constant equivalent potential temperature (θ_e).



A complete Skew-T Log-P diagram, used to visualize changes in the atmosphere with altitude. ([CC BY-NC-SA 4.0](https://creativecommons.org/licenses/by-nc-sa/4.0/)).

Here is a complete Skew-T Log-P diagram. All of the lines look confusing and complicated when combined, but each represents a constant change in one variable.

Let's look at another real balloon sounding. This time launched from Hilo during Hurricane Lane.



Balloon sounding launched from Hilo as Hurricane Lane impacted the Big Island. The sounding was retrieved from the Upper Air Soundings portion of the University of Wyoming Weather Web: <http://weather.uwyo.edu/upperair/sounding.html>. (Copyright 2018 by University of Wyoming Department of Atmospheric Science, used with permission.)

On this Skew-T diagram, all of the same lines are there. Horizontal blue lines are isobars, slanted blue lines are isotherms, slanted purple lines are isohumes, the green lines are the dry adiabats, and the blue curved lines are the moist adiabats. The T (right) and T_d (left) black lines are close together and sometimes overlap in the lowest 500 hPa of the atmosphere because the lower levels are incredibly moist, and a deep cloud layer extended up to nearly 6 km altitude.

Finding the Lifting Condensation Level (LCL)

When plotting a sounding on a Skew-T diagram, you may have a selection of data similar to the example given below. You will likely have pressure, temperature (T), and a dew point temperature (T_d) with altitude.

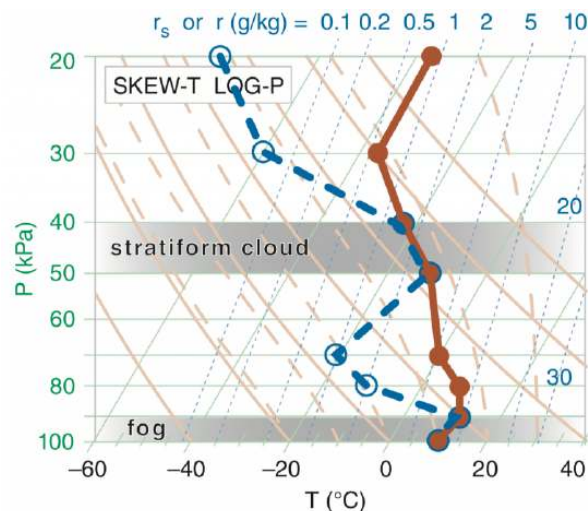
Sample Application

Plot the following data on a skew-T log-P diagram.

P (kPa)	T (°C)	T_d (°C)
20	-36	-80
30	-36	-60
40	-23	-23
50	-12	-12
70	0	-20
80	8	-10
90	10	10
100	10	10

Sample atmospheric data to be plotted on the skew-T diagrams ([CC BY-NC-SA 4.0](https://creativecommons.org/licenses/by-nc-sa/4.0/)).

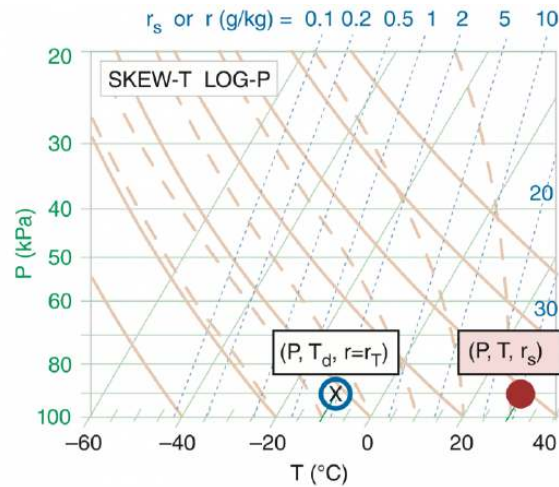
In order to plot the sounding, it is easiest to start by finding the pressure level and then move to the right to plot the temperature and dew point temperature. Pay careful attention to the fact that the isotherms are skewed. Rotate the axis in your mind when you plot your temperature and dew point. Once you have plotted all of your temperatures and dew points, you will have a vertical temperature and humidity profile of the atmosphere.



Sample example plotted ([CC BY-NC-SA 4.0](https://creativecommons.org/licenses/by-nc-sa/4.0/)).

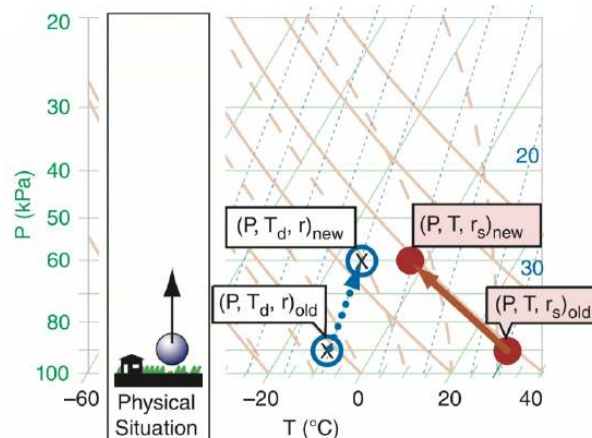
Now that we plotted the sounding, it is useful to know how a rising air parcel will behave when placed in this environment. Is the atmosphere stable, unstable, or conditionally unstable? We can determine this by estimating the rate at which a rising parcel will cool and drawing a parcel path upward. A rising air parcel will cool at the dry adiabatic lapse rate until it is saturated, after which it will cool at the moist adiabatic lapse rate. How do we know when a parcel will be saturated? First we need to find the **Lifting Condensation Level (LCL)**.

The **Lifting Condensation Level (LCL)** is the level at which the water vapor in an air parcel that is lifted dry adiabatically will be saturated.

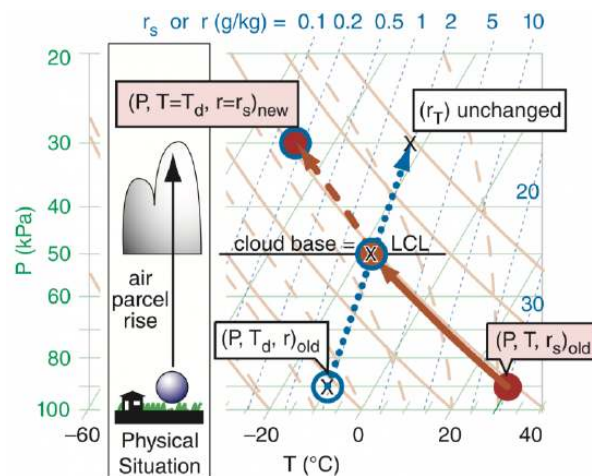


The red dot is air temperature and the blue circle is dew point temperature. This diagram is an example of an unsaturated air parcel. Stull Figure 5.7 ([CC BY-NC-SA 4.0](#)).

To find the LCL, start at the surface (or the pressure level closest to the surface, typically 1000 hPa) and plot the temperature and dewpoint temperature. In the case of the example above, the surface pressure level must be at a raised elevation with $P_{\text{surf}} = 90 \text{ kPa}$ or 900 hPa, $T = 30^\circ\text{C}$, and $T_d = -10^\circ\text{C}$. Imagine that the air parcel has the same temperature and dewpoint temperature as the environment at first. Initially, it will cool at the dry adiabatic lapse rate as it rises. First, follow the surface temperature upward along a dry adiabat. In all likelihood, the temperature will not be directly along a marked dry adiabat line as it is in the example so follow a line upward parallel to a dry adiabat. Similarly, start at your surface dew point and follow the isohume (constant mixing ratio line) upward because the moisture content of the air parcel does not change with dry lifting. Draw these lines upward until they intersect. This intersection will give you the level of the lifting condensation level (LCL).



Follow the dry adiabat and isohume lines until they intersect ([CC BY-NC-SA 4.0](#)).



The place where the two lines intersect is the lifting condensation level ([CC BY-NC-SA 4.0](#)).

In this example, the surface temperature and dewpoint temperature line up nicely with an isohume and a dry adiabat line, but this typically won't be the case with a real sounding. The procedure, however, will be the same. The LCL marks the approximate cloud base height for convective clouds (cumulus type), where rising air first becomes saturated.

After the air parcel has been lifted dry adiabatically to the LCL, it becomes saturated. As we know, a saturated air parcel cools at the smaller *moist adiabatic lapse rate*. From the LCL, follow a line parallel to a moist adiabat upward to get the approximate lapse rate of your parcel as it rises. In the example soundings from Hilo and Lihue shown earlier, this same line is plotted in a light grey color from the surface all the way up in the atmosphere. It shows the temperature a surface based parcel would have when lifted through the troposphere.

As you follow an air parcel temperature upward moist adiabatically, the point at which it intersects the environmental temperature profile (where your parcel becomes warmer than its environment) is called the **Level of Free Convection**, or the **LFC**.

As you continue following the air parcel path upward moist adiabatically from the LFC, the point where it intersects the sounding again (the point where your parcel becomes cooler than its environment) is called the **Equilibrium Level (EL)**.

Normand's Rule for Wet-bulb Temperature

You can estimate the surface wet bulb temperature by taking the LCL example one step further. *Normand's Rule* is used to calculate the wet-bulb temperature from the air temperature and the dew point temperature. The wet bulb temperature is always between the dew point and the dry bulb temperature ($T_d \leq T_w \leq T$). To find the wet bulb temperature on a Skew-T Log-P diagram, follow the surface T upwards along a dry adiabat, and the surface T_d upwards along a isohume. Where they meet is the LCL, as just explained. Next, follow a moist adiabat back down to the surface. Where the moist adiabat intersects the surface is the wet-bulb temperature value.

5.3: Skew-T Log-P Diagram is shared under a [CC BY-SA 4.0](#) license and was authored, remixed, and/or curated by LibreTexts.

5.4: CAPE and CIN

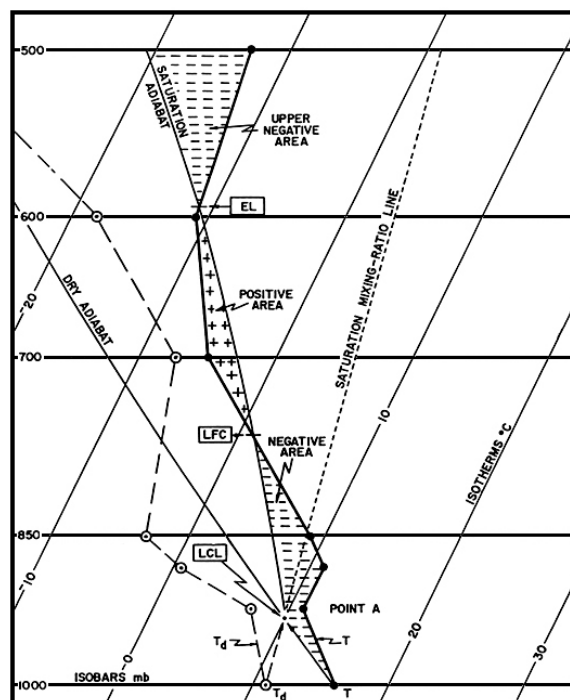
The “positive area” between the parcel path and the environmental temperature profile, traced out between the LFC and the EL (where the parcel is warmer than the environment) gives a measure of the **Convective Available Potential Energy**, or **CAPE**, given in units of $\text{J}\cdot\text{kg}^{-1}$. This is an estimate of the buoyant energy of a parcel and can provide a means of estimating the strength of any convection that may occur. CAPE can also provide an estimate of the maximum updraft intensity in a thunderstorm.

$$w_{\max} \sim 0.60 \cdot (2 \cdot \text{CAPE})^{\frac{1}{2}}$$

$w_{\max}$ is the estimated maximum vertical motion as a result of CAPE.

Convective Inhibition, or **CIN** is essentially negative CAPE, also in $\text{J}\cdot\text{kg}^{-1}$. It is the negative area between the parcel path and the environmental temperature curve where the parcel is cooler than the environment. The larger the value of CIN, the greater the negative buoyant energy that acts against CAPE. CIN sometimes acts as a “cap” on convection. If you have large CAPE but also large CIN, your CAPE may not be fully realized as buoyant energy and you may not have any convection. However, if your parcel is able to break through the cap, that is, if it is able to rise and become warmer than the environment, convection may be strong.

The figure below shows the locations of the LFC and EL, and shades in both positive and negative areas between the parcel path and the environmental temperature profile.

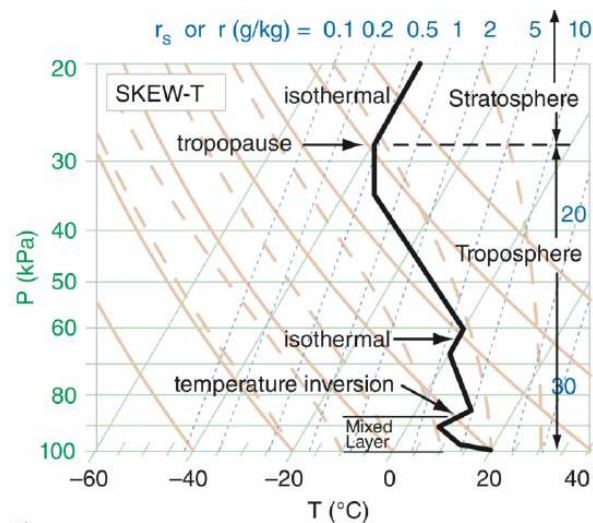


The locations of the LFC and EL in the vertical sounding are shown, found as the positive and negative areas between the parcel path and the environmental temperature profile ([Public Domain](#)).

In the Lihue and Hilo soundings shown previously, values of CAPE and CIN are given in $\text{J}\cdot\text{kg}^{-1}$ in the column on the right hand side. Note that CIN is written as “CINS” and denoted as a negative value.

Locating the Tropopause

Recall that the standard temperature decreases with height within the troposphere, but becomes isothermal with height within the tropopause, and increases with height in the stratosphere. With this knowledge, the location of the tropopause, given by its pressure level, can be determined by examining a plotted sounding. In the upper part of your sounding, look for where the temperature profile becomes isothermal (parallel to your skewed isotherms) or for an inversion (where the temperature increases with height, which will be tilted to the right more than your isotherms). The base of the isothermal layer in your sounding is the tropopause.



A plotted sounding with two isothermal layers and a temperature inversion denoted ([CC BY-NC-SA 4.0](#)).

There are many things we can learn about the atmosphere from Skew-T Log-P diagrams. Here we've provided just the basics to get you started.

Chapter 5: Questions to Consider

- 1.
- 2.
3. Drag and drop terms to their correct position in the atmospheric stability diagram below:
4. What does the Lifting Condensation Level (LCL) represent? How can it be found on a Skew-T diagram?
5. What is CAPE? How can it be found on a Skew-T diagram?

Selected Practice Question Answers:

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CHAPTER OVERVIEW

6: Clouds

Alison Nugent and Shintaro Russell

Learning Objectives

By the end of this chapter, you should be able to:

- Define what a “cloud” is
- Describe the importance of cloud condensation nuclei in cloud formation
- Describe the difference between cumuliform and stratiform cloud classes
- Differentiate between and identify major cloud types including cumulus, stratus, cirrus, altostratus, etc.
- In addition to identifying the cloud, you should know the general altitude (low, mid, or high), cloud class (convective or stratiform) and whether or not it is a rain cloud
- Discuss when/where/why clouds form

[6.1: Introduction](#)

[6.2: Cloud Condensation Nuclei](#)

[6.3: Cloud Naming Conventions](#)

[6.4: Cloud Identification Examples](#)

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6.1: Introduction

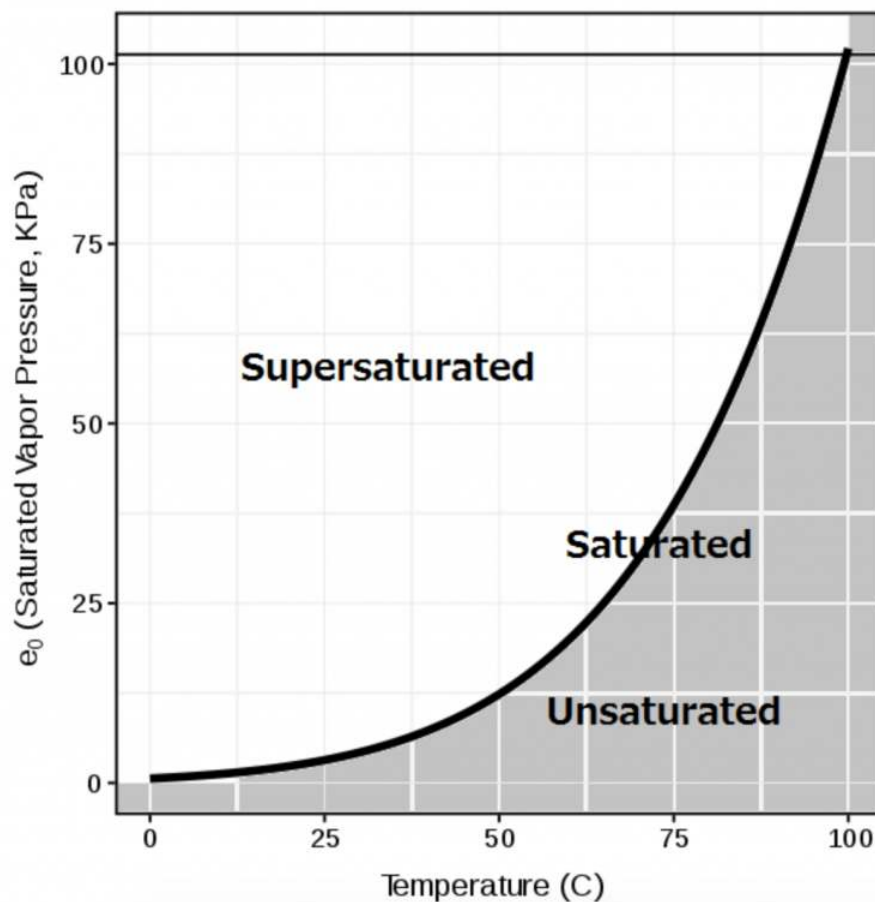


Cumulus clouds (Photo by Shintaro Russell).

What is a cloud and how does it form?

A cloud is a collection of suspended particles of water droplets and/or ice crystals in the atmosphere. The cloud droplets or crystals are so small that their terminal velocity, the highest velocity possible as an object falls through a fluid, is negligible. They are falling but they fall so slowly that they appear to be suspended in the air.

Recall from Chapter 4, the Clausius-Clapeyron diagram. The line represents the vapor pressure at saturation for a given temperature. The region to the right of the line represents air that is sub-saturated, and the region to the left of the line represents air that is super-saturated. Sub-saturated air has a relative humidity below 100% and super-saturated air has a relative humidity above 100%.



The Clausius-Clapeyron relationship showing that saturated vapor pressure vs. temperature have an exponential relationship ([CC BY-SA 4.0](#)).

Imagine a point on the right side of the line, at $T = 75^{\circ}\text{C}$, and $e_0 = 12.5\text{kPa}$. To become saturated, the air either needs to cool with the same moisture content (move left on a constant e_0 line), or increase the moisture content (move upward on a constant T line) until the air meets the saturation curve. To put this another way, we know that clouds form when the atmosphere is saturated. The list below describes the three ways air parcels can become saturated.

Clouds can form from

1. Adding Moisture (or mixing with cool moist air)
 - Sea smoke
 - Contrails
 - Cooling towers
 - Exhaling
2. Cooling by Removing Heat
 - Radiation fog: clear nights on land
 - Advection fog: air moving over cold ocean currents
3. Cooling by Adiabatic Expansion
 - Upward air motion
 - Vortices, like wing tip vortices on aircraft, or tornados
 - Supersonic flight

Saturation is typically achieved by either adding moisture until the dew point temperature is equivalent to the temperature or cooling until the temperature is lowered to the dew point temperature. In some cases, both moisturizing and cooling can happen at

the same time. The mixture of two unsaturated air parcels can even cause saturation in the resulting mixture. For example, a person's breath on a cold day or jet contrails can both form clouds because of this mixing process.

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6.2: Cloud Condensation Nuclei

When air becomes saturated, excess water vapor in the atmosphere can condense to form liquid water. However, water vapor requires a surface on which to condense. The surface is provided by tiny particulates in the atmosphere known as aerosols, or more generally, cloud condensation nuclei (CCN). Aerosols are everywhere in the atmosphere. They can be composed of earth matter like dust, clay, or soot. They can also be composed of sea salt from the ocean, black carbon from fires, or sulfate from volcanic activity. They can be composed of material from plant or organism life, like sulfates and nitrates or volatile organic compounds, or even phytoplankton. Anything that is small enough to be lofted and suspended in the atmosphere has the potential to be an aerosol. An aerosol that is hygroscopic, or “water-liking”, has the potential to be a cloud condensation nuclei.

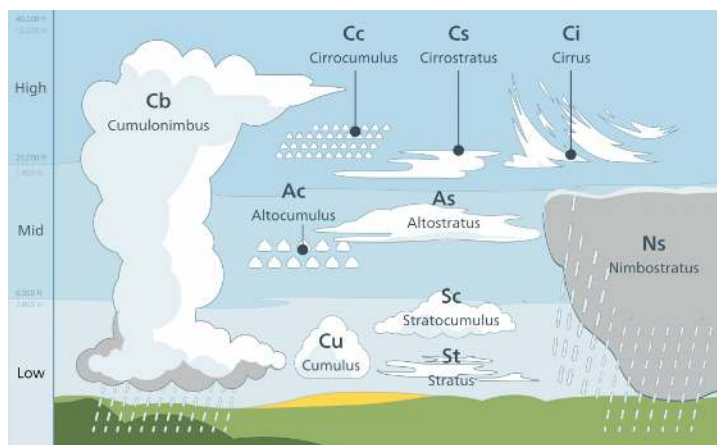
If there were no aerosols in the atmosphere, relative humidities well over 100% would be needed for clouds to form. In laboratory studies with a clean atmosphere (with no aerosols) relative humidities up to 400% have been measured. In Earth’s atmosphere, aerosols are abundant, so water vapor does not struggle to find CCN to condense onto.

Another type of aerosol that helps with the formation of cold clouds, or clouds composed of ice crystals, are called Ice Nuclei (IN). Dust is a particularly good ice nuclei. Materials with a hexagonal shape, like ice crystals, also make particularly good surfaces for ice to form on. A common one used in cloud seeding is called silver iodide, AgI. We’ll discuss this more in the following chapter on precipitation.

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6.3: Cloud Naming Conventions

Once a cloud forms, how do we know what to call it? Clouds form in many environments and look different depending on those environments, and depending on whether they're composed of liquid droplets or ice crystals. The following diagram gives a brief overview of the many cloud types, along with their common abbreviations.



This diagram shows the common cloud types and associated altitudes. The cloud naming convention involves dividing the atmosphere into Low, Middle, and High level clouds, and typically cumuliiform and stratiform clouds ([CC BY-SA 3.0](https://creativecommons.org/licenses/by-sa/3.0/)).

Clouds can be identified based on their appearance, shape, and altitude. The two primary cloud categories are cumuliiform and stratiform clouds.

By Appearance and Shape

Cumuliiform clouds develop as a result of vertical motions by atmospheric instability. They are *convective* clouds meaning that they form in air parcels that are buoyant and are undergoing convection, which is the transfer of heat or mixing within a fluid due to warm air rising and cool air sinking. Examples of cumuliiform clouds include cumulus, cumulus congestus, and cumulonimbus.

Stratiform clouds are horizontally layered clouds. They tend to spread into wide regions, and take on an appearance of a sheet or blanket. They typically form when a layer of air is brought to saturation but is thermodynamically stable, or when a convective cloud meets a stable layer and spreads out in a layered fashion. Examples of stratiform clouds include nimbostratus, stratus, altostratus, cirrostratus, and cirrus.

By Water Phase

Clouds composed of only liquid water droplets are called “warm clouds”, and typically have clearly defined edges. Low altitude clouds are usually warm clouds.

Clouds made up of only ice crystals are called “cold clouds” and typically have fuzzy looking edges. The edges look fuzzy and not well defined because it takes longer to go from the ice phase to the vapor phase as compared to warm clouds. This transition time scale results in a larger cloud to dry air transition region. High altitude clouds are always cold clouds.

Clouds composed of liquid water and ice crystals are called “mixed phase clouds”. It is difficult to distinguish by eye whether a cloud is mixed phase. Often the decision comes down to the height of the cloud, knowing that clouds that extend from the surface up high in the atmosphere likely have a mixture of both liquid droplets and ice crystals.

By Altitude

Clouds at a high altitude have the prefix “cirro” or “cirrus”. Due to the high altitude, the cirrus and cirrostratus are made out of ice crystals.

Clouds at mid-altitude have the prefix “alto”. They are usually made out of liquid droplets, but can be a mixture between liquid droplets and ice crystals.

Low level clouds don’t have a particular prefix.

By Characteristics

The prefix “nimbo” or suffix “nimbus” indicates a precipitating cloud. Nimbostratus usually have light to moderate precipitation, whereas cumulonimbus or thunderstorm clouds have heavy precipitation and sometimes even hail.

Other

The above clouds represent the primary cloud types, but many other cloud names and cloud types are used in atmospheric sciences. A few of the more common cloud names are given below.

Lenticular clouds are stationary lens-shaped clouds with a smooth appearance that usually forms over the summit of a mountain or on the lee wave crest. They are also referred to as lee-wave clouds or mountain-wave clouds.

Mammatus clouds are formed from downward convection, typically in the anvil portion of a cumulonimbus. They have a distinctive look that makes them easy to spot, especially at sunset. They are made up of hanging pouches that look a little bit like udders.

Fog

A cloud that forms at or near the ground is called fog. The main types of fog are upslope, radiation, advection, precipitation or frontal, and steam fog.

Upslope, radiation, and advection fogs develop *due to cooling*. Upslope fog develops as stable, moist air rises over topography like a hill or mountain. Radiation fog occurs over land as radiational cooling lowers the air temperature to its dew point temperature. Radiation fog usually occurs during calm, clear nights. Advection fog forms as warm, moist air flows over a colder surface and the air cools to its dew point temperature.

Frontal and steam fogs are formed by the *addition of water vapor*. Frontal fog develops when warm raindrops evaporate in a colder airmass. Steam fog or sea smoke forms as cold air moves over warmer water.

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6.4: Cloud Identification Examples

The image gallery below contains many different clouds types. The clouds are labelled and described in their figure captions.



Stratus clouds over Diamond Head. Stratus clouds are a low level cloud type, composed of liquid droplets. They form in an atmosphere that is stably stratified, hence the featureless clouds. (Photo by Shintaro Russell)



Cumulus clouds are a convective low level cloud type. They're composed of liquid droplets and form in an unstable, or conditionally unstable atmosphere. (Photo by Shintaro Russell)



Cirrus clouds are a high level cloud type. Cirrus clouds are composed of tiny ice crystals. (Photo by Shintaro Russell)



Cumulonimbus and Altocumulus clouds. The cumulonimbus cloud is the cloud that has a large vertical extent in the center of the image. The altocumulus clouds are on the top and the top right of the image. Cumulonimbus clouds have the suffix “nimbus” because they’re raining, and because they extend through so many levels of the atmosphere, they’re typically mixed phase, having both liquid and ice inside. (Photo by Shintaro Russell)



Cirrocumulus clouds are high level convective clouds. They're typically constrained to a thin layer. (Photo by Shintaro Russell)



As cumulus clouds begin to grow vertically, they're called cumulus congestus like these convective clouds in the image. They're not cumulonimbus yet, but they're deeper than the average cumulus. (Photo by Shintaro Russell)



Altocumulus clouds are mid-level cumulus clouds. Altocumulus and cirrocumulus are often difficult to differentiate from a photograph because the altitude is hard to gauge. However, these clouds look more well defined than the cirrocumulus in the prior image, meaning they're likely made of liquid droplets rather than ice crystals. (Photo by Shintaro Russell)



Stratocumulus clouds are low level layered clouds. They're called strato – meaning layered, and cumulus – meaning convective because they're convective-looking, but confined to a layer. (Photo by Shintaro Russell)

A link to an entire cloud gallery put together by Shintaro Russell is included [here](#). All of the photos were taken here in Hawaii, and so many cloud types are represented! Enjoy!

Chapter 6: Questions to Consider

1. What relationship is shown in the Clausius-Clapeyron diagram?
2. Why are CCN important?
3. What is the difference between cumuliform and stratiform clouds?
4. Drag and drop the cloud type to its correct position in the diagram below:

Selected Practice Question Answers:

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CHAPTER OVERVIEW

7: Precipitation Processes

Alison Nugent and David DeCou

Learning Objectives

By the end of this chapter, you should be able to:

- Recall the importance of cloud condensation nuclei and aerosols
- Calculate the speed of a falling cloud droplet and raindrop
- Describe the Collision-Coalescence process
- Describe the Ice Phase (Wegener-Bergeron-Findeisen) process

[7.1: Introduction](#)

[7.2: Cloud Droplets](#)

[7.3: Precipitation Types](#)

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7.1: Introduction



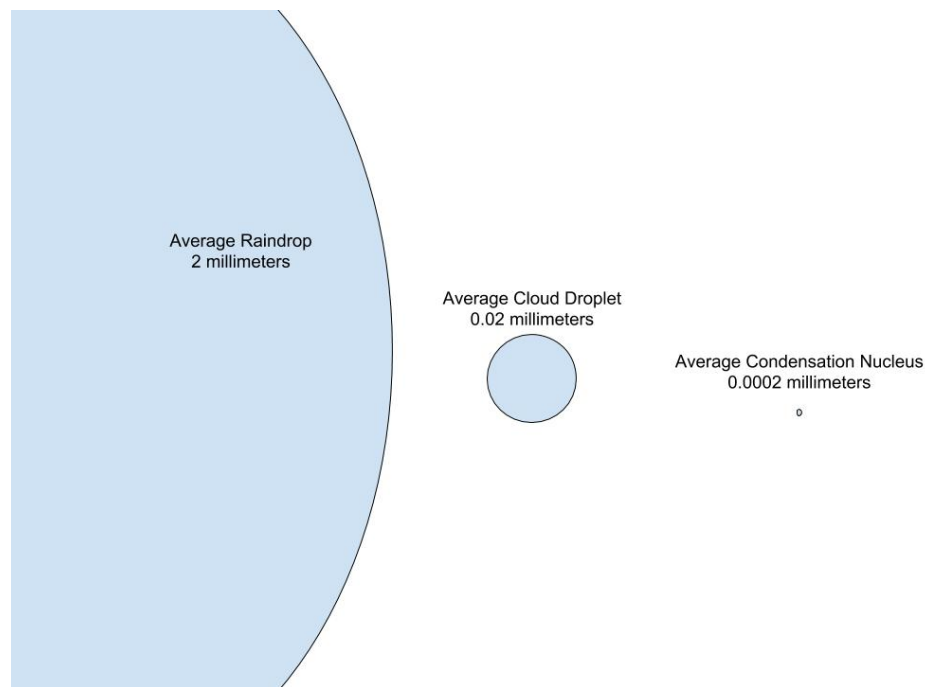
Raindrops falling on a body of water ([CC 0](#)).

Sometimes rain feels like a gentle mist but at other times its a heavy downpour that floods streets and sidewalks. Many times, clouds cover the skies but never produce any precipitation at all. This leads us to question: why does it rain and do raindrop sizes vary? What is the relationship between raindrops and cloud droplets, and by what processes do each form? You know that clouds form by condensation but, apparently, condensation by itself is a necessary but insufficient condition for rain. We will explore why this is by examining cloud droplets and raindrops in more detail.

The average cloud droplet is very small with an average diameter of about 20 micrometers (μm), which is the same as $20 \times 10^{-6}\text{m}$, 0.002 cm, or 0.02 mm. This diameter is about 100 times smaller than your average raindrop.

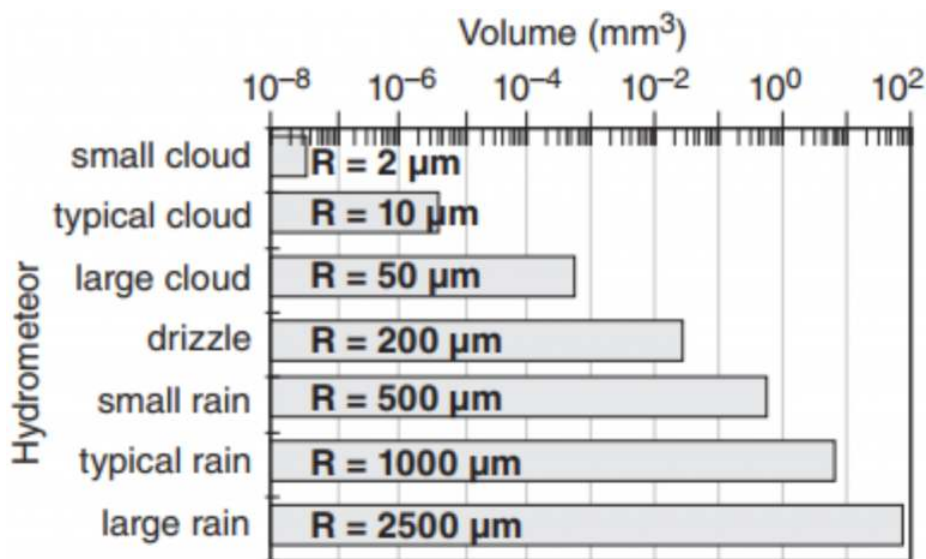
Pro Tip: 1 micron (μm) is the same as one-millionth of a meter ($1 \times 10^{-6}\text{m}$). In cloud microphysics, microns are the standard scale of measure.

The following image gives a sense of the difference in scale between raindrops (left), cloud droplets (center), and cloud condensation nuclei (right). The average raindrop has a diameter of 2 mm, and the average condensation nucleus has a diameter around 0.0002 mm.



Comparison of raindrop, cloud droplet, and condensation nucleus sizes, given as diameter in mm (Image Created by Britt Seifert).

When considering the volume of the droplets or particles, this differences quickly grows. The following image shows the volume of various cloud droplets and rain drops on a log scale.



Type of hydrometeor vs. radius “R”, given in microns, and drop volume (mm^3) on a log scale ([CC BY-NC-SA](#)).

Notice how cloud droplet sizes range from $2 \mu\text{m}$ to $50 \mu\text{m}$ and raindrop sizes range from $200 \mu\text{m}$ to $2500 \mu\text{m}$. Liquid drops exist on a size spectrum from about $1 \mu\text{m}$ to almost $5,000 \mu\text{m}$ (or 0.5 cm). The minimum size for a cloud droplet is effectively set by the surface tension required to keep the H_2O molecules together. The smaller the droplet, the higher the surface tension necessary. The maximum size for a raindrop is limited by drop breakup because when the drop becomes too large, air friction will break it up into a bunch of smaller droplets.

In general, the only difference between a cloud droplet and a raindrop is that a raindrop has a non-negligible fall velocity. On a continuous spectrum of sizes, at some point the gravitational pull on water drops in the atmosphere becomes large enough not to ignore. While all drops will fall, the larger the drops are, the faster they fall.

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7.2: Cloud Droplets

Recall from the previous chapter on clouds that **cloud condensation nuclei** (CCNs) are required for water vapor to condense onto. Many CCNs are hygroscopic, meaning they tend to absorb moisture, so condensation may start before the relative humidity reaches 100%. For example, when condensation occurs on salt particles, which are extremely hygroscopic, condensation can begin at 80% relative humidity or lower.

Imagine there are many CCNs of differing sizes in a body of humid, but unsaturated air. If the air were to be lifted by a mountain or in a rising thermal, it would cool, and the relative humidity would increase. As the air nears saturation, condensation will begin to occur on the largest and most hygroscopic CCN. At some later time, a cloud of many small cloud droplets, far too small to fall at any significant speed, will form.

The terminal velocity of a falling cloud droplet (with radius “ r ” less than 40 μm) is given by the following equation from Stoke’s Drag Law:

$$w_{T, \text{cloud}} = (-1.19 * 10^8 m^{-1} s^{-1}) * r^2$$

where “ r ” needs to be expressed in meters. A simple calculation will show that it takes hours, if not days, for a cloud droplet to fall from even low altitudes. The friction provided by the air or even tiny upward air currents will keep cloud droplets suspended for long periods.

Terminal velocity implies that a steady state has been reached in the fall velocity with a balance between the downward gravitational force on the drops and the upward frictional drag of air.

Raindrops

For raindrops, a different equation is used to approximate the fall speed. For spherical raindrops,

$$w_{T, \text{rain}} = (-220 m^{1/2} s^{-1}) * \left(\frac{\rho_0}{\rho_{\text{air}}} * r \right)^{1/2}$$

where ρ_0 is a reference value of density, typically 1.2 kg m^{-3} and ρ_{air} is the density of air where the raindrop exists. Again, “ r ” is the radius of the drop, given in meters. High up in the atmosphere when ρ_{air} is small, the speed of a falling raindrop, $w_{T, \text{rain}}$, will be faster than near the surface when ρ_{air} is similar magnitude to ρ_0 . As the air density increases, the frictional drag on a drop also increases.

Note that this equation for raindrops is a vast simplification because raindrops are not typically spherical shaped. As they fall, the passing air deforms them into pancake shaped drops. However, this equation provides an approximation for fall velocity.

So how can cloud droplets grow to form raindrops? The condensation process is not enough and is far too slow to produce raindrops. From a volumetric view, it takes 1 million cloud droplets (10 μm radius) to combine together to make one single raindrop (1000 μm = 1 mm radius). There has to be another faster process by which cloud droplets can grow or combine together to become large and heavy enough to fall.

We will discuss two primary rain formation theories.

1. Collision-Coalescence Process
2. Ice Phase Processes (Wegener-Bergeron-Findeisen)

Collision-Coalescence Process

In warm clouds, where all of the cloud droplets are liquid, the **collision-coalescence** process is the primary mechanism responsible for producing precipitation. This is thought to be the case especially over tropical oceans. The collision-coalescence process is exactly as it sounds: cloud droplets collide and coalesce or stick together. Larger cloud droplets have slightly higher terminal velocities, because they have a smaller surface-area-to-weight ratios. This advantage allows them to fall faster and collide with smaller cloud droplets. Sometimes the cloud droplets will stick together and coalesce to form a larger droplet. This begins a positive feedback where these larger droplets then fall even faster, collide with even more smaller droplets in their path, and aggregate more and more cloud droplets together. However, note that collision between cloud droplets does not always mean that

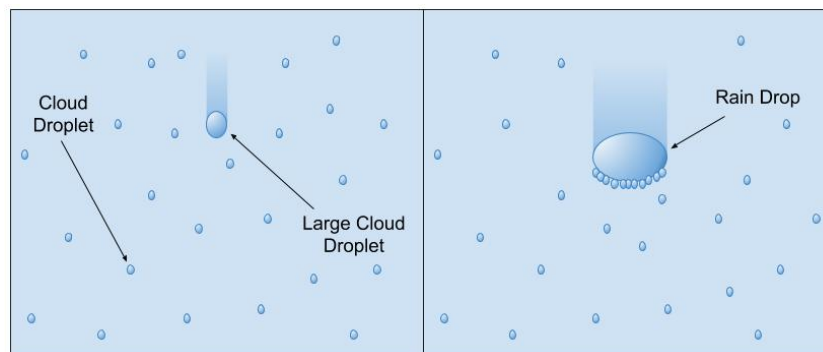
coalescence will occur. Sometimes droplets will bounce apart during collision if their surface tensions are too strong. For collision-coalescence to begin, a cloud needs to have a wide distribution of cloud droplet sizes. This can occur from a variation in CCN type—for example, sea salt aerosols are particularly large—or from random collisions between droplets.

The total amount of liquid water in a cloud as well as the time that a cloud droplet spends inside of a cloud influences how large it can grow through the collision-coalescence process. The cloud height is of course a factor here, but its a little more complicated than that. Rising motion in a cloud will slow the downward speed of a falling droplet. This can act to increase the amount of a time a cloud droplet spends in a cloud, as well as the size it will grow. Let's cover a few examples.

Deep cumulus clouds with convective updrafts tend to produce larger raindrops because upward motion is strong and droplets have a long time in the cloud to grow. In fact, the droplets need to become sufficiently large in order for their fall velocity to overcome the updraft velocity.

On the other hand, stratus clouds are typically not very thick and have weak updrafts, so droplets in these clouds don't spend a long time in the cloud itself, and therefore are not be able to grow very large. If there is moist air below the stratus cloud, the drops may reach the ground as a light drizzle. However, if there is dry air below the stratus cloud, the drops may evaporate before they are able to reach the ground.

To summarize, in warm clouds, cloud droplets grow to precipitation sized drops through the collision-coalescence process. The most important factor in raindrop production is the liquid water content of a cloud. Assuming the cloud has sufficient water, other factors that affect raindrop production are: thickness of the cloud; strength of updrafts within the cloud; cloud drop distribution of sizes; and difference in electric charge of the droplets and the cloud itself. Thin stratus clouds with weak vertical motion may produce weak drizzle, if any, while tall cumulus clouds with strong updrafts can produce heavy rain showers. The following image illustrates the collision-coalescence process of raindrop production.



The collision-coalescence process occurring in warm clouds. The left image shows the importance of a range of cloud droplet sizes to initiate the collision-coalescence process while the image at right shows the acceleration of the process once a raindrop forms (Image Created by Britt Seifert).

Ice Phase Process

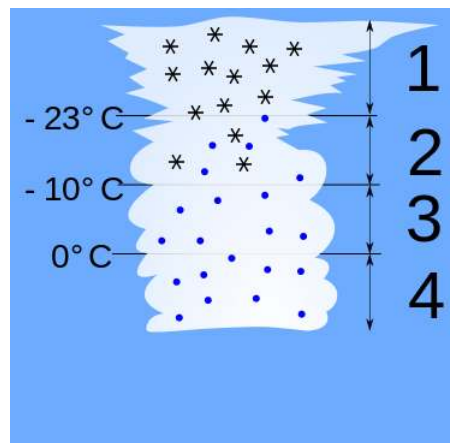


Ice crystals forming on a window at 30,000 feet ([CC BY-SA 3.0](#)).

Outside of the tropics, the ice phase process of rain formation is the primary mechanism producing most of the world's precipitation. The ice phase process occurs in cold clouds or clouds with temperatures below 0°C . To understand why, we need to know something about freezing of liquid water droplets.

Supercooled Water and Ice Nuclei

Supercooled water is liquid water that exists below the freezing point of water (below 0°C). Similar to how cloud droplets need a surface on which to condense, ice crystals also need a nucleus or ice embryo to freeze. Without an ice nucleus, liquid water drops can remain liquid in temperatures as low as -40°C . Once beyond -40°C , all hydrometeors (water particles) will exist in the solid state. Typically the distribution of liquid and solid hydrometeors in a cloud looks like the following image.



A simple diagram showing the distribution of liquid and solid phase hydrometeors in a mixed phase cloud ([CC BY-SA 3.0](#)).

At low elevations above freezing (region 4), the hydrometeors in the cloud exist as liquid droplets. Above the freezing level (region 3), supercooled liquid droplets exist. Above that, some liquid droplets begin to freeze, and both liquid and ice phase hydrometeors co-exist (region 2). Finally, above some level where the temperature is cold enough, all hydrometeors will exist in their solid state (region 1).

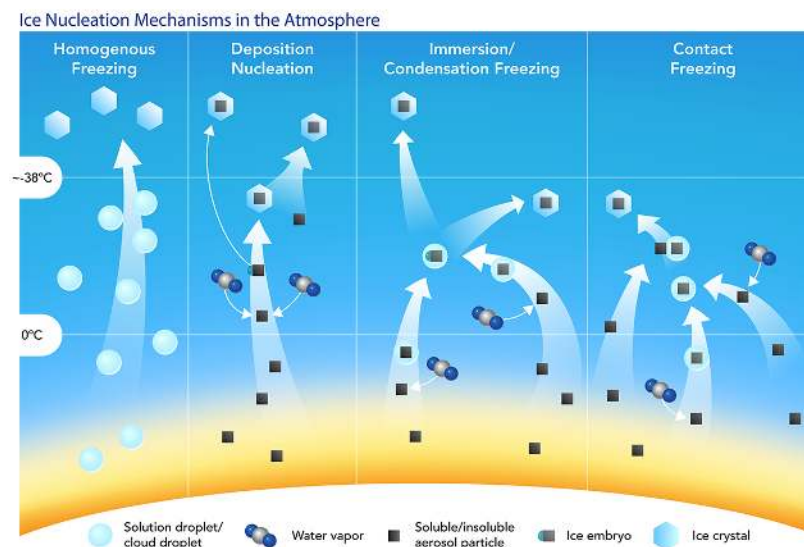
When liquid water droplets freeze without any sort of nucleus, this is known as **homogenous** or **spontaneous freezing**. While this occurs within a large body of freshwater at temperatures slightly below 0°C , cloud droplets will not freeze spontaneously until temperatures are -40°C or lower.

For droplets to freeze spontaneously, enough molecules within the droplet must form a rigid pattern and become a tiny ice structure known as an *ice embryo*. When this embryo grows sufficiently large, at a certain size it will act as an ice nucleus, which are described below. The other molecules in the droplet then become attached to the ice structure and the entire droplet freezes.

Tiny ice embryos are able to form when water drops just below freezing, but typically at these temperatures there is enough thermal agitation to weaken their structure and break them apart. At lower temperatures, there is less thermal motion, and ice embryos have a better chance of growing large enough to freeze the surrounding water. When you have larger volumes of water, ice embryos have a better chance of growing large enough to freeze the surrounding liquid before being broken up, but this becomes more and more difficult with smaller volumes of water. Only the largest cloud droplets can freeze spontaneously without a nucleus at temperatures below -40°C . In most cases, ice nuclei are required for ice crystals to form in sub-freezing clouds.

Just as CCNs are required for liquid cloud droplets to form, ice crystals form on particles called **ice nuclei (IN)**. Particles serve as effective IN if they have similar geometry to an ice crystal, for example, ice itself is an effective IN. There are not many IN in the atmosphere, especially at temperatures above -10°C , but certain types of particles become active IN with lower temperatures. For example, dust can be an effective IN. Ice nuclei are rare compared to hygroscopic cloud condensation nuclei.

Some IN allow water vapor to immediately become solid ice when they come in contact together. These are known as *deposition nuclei* because the water vapor changes phase directly into solid ice without becoming liquid first (phase change from gas to solid is called deposition). IN that are effective at causing the freezing of supercooled liquid droplets are called **freezing nuclei**. Some freezing nuclei must be immersed in a liquid drop in order to freeze it, while others are effective at inducing condensation and then freezing. Many freezing nuclei will cause supercooled droplets to freeze as soon as they collide, which is called **contact freezing**, and these nuclei are referred to as **contact nuclei**. These different freezing methods are outlined in the figure below.



Four primary mechanisms that form ice in the atmosphere: homogeneous freezing; deposition nucleation; immersion freezing; and contact freezing ([CC BY-SA 4.0](#)).

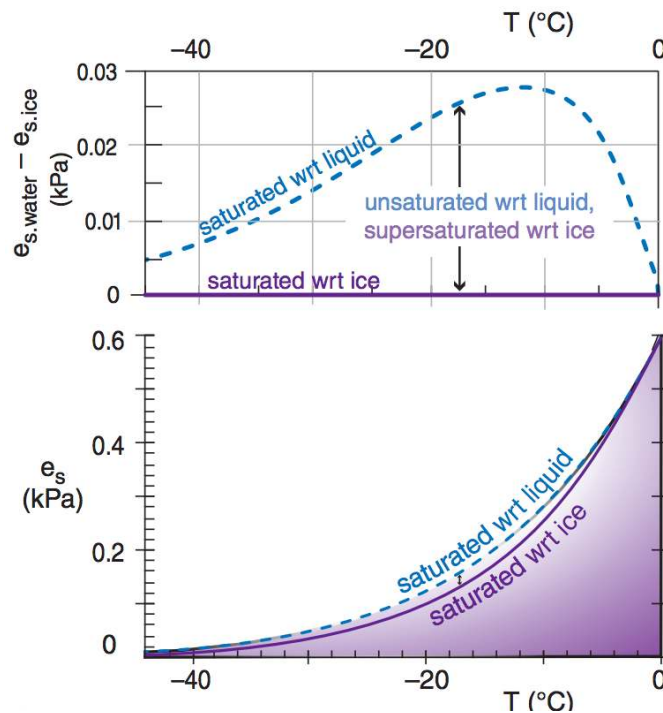
To summarize, cloud droplets may freeze spontaneously, but only at very low temperatures. Ice nuclei can help the growth of ice crystals, but they are not naturally abundant.

Saturation Vapor Pressure

So, we have cold clouds that contain many more liquid cloud droplets than ice crystals, even at sub-freezing temperatures, and these particles are not large/heavy enough to precipitate out of the cloud. How do we get rain and snow out of the ice-crystal process then?

Imagine a cloud with super cooled liquid water and saturated air. At saturation, the liquid droplets are in equilibrium with the water vapor in the air. The number of water molecules leaving and entering the surface of the liquid droplets are equivalent. Now imagine that an ice crystal forms by one of the processes described above. In the below-freezing portion of a cloud, this ice crystal is

surrounded by many liquid supercooled droplets. Because the saturation vapor pressures with respect to liquid and ice are slightly different, the presence of this new ice crystal has a big impact on the cloud.



The saturation vapor pressure with respect to liquid (blue) and with respect to ice (purple). The difference is shown at the top ([CC BY-NC-SA](#)).

With respect to liquid, the liquid droplets were at saturation. But now, with respect to ice, the ice crystal is in an environment that is supersaturated. You can think of this as the following: it is easier for water molecules to escape a liquid surface through evaporation than to escape a solid surface. This means that there will be many more molecules escaping the liquid water surface at a given temperature and will require more water vapor around it in order to keep the droplet in saturation. At the same temperature, the saturation vapor pressure above a surface of water is greater than the saturation vapor pressure above a surface of ice.

This difference in saturation vapor pressure with respect to water and ice causes water vapor molecules to deposit from the environment onto the ice crystal. Because the vapor molecules are being removed from the environment around the liquid droplet, the vapor pressure with respect to the water surface decreases. This throws the water droplets out of equilibrium with their surroundings, causing them to evaporate, replenishing the removed water vapor from the environment. This provides additional moisture for the ice crystal, allowing it to grow at the expense of the liquid droplets.

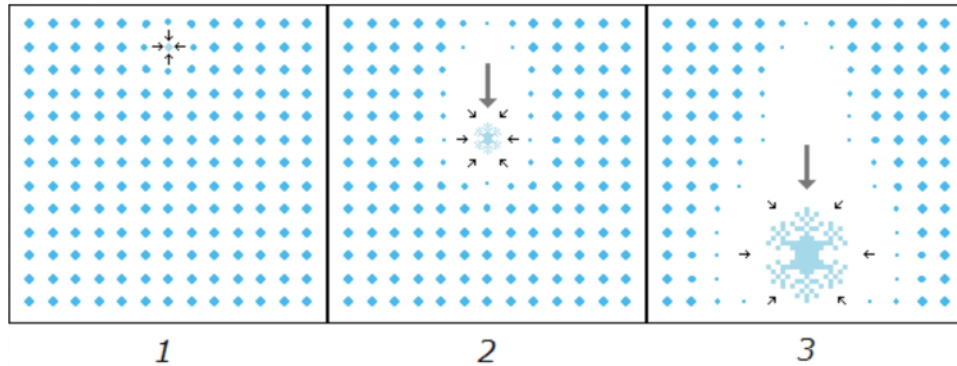
This process is called the **Wegener-Bergeron-Findeisen** (or more generally, the ice phase) process. Ice crystals in a sub-freezing region of a cloud will grow larger at the expense of surrounding water droplets.

Falling Ice Crystals

Imagine this happening throughout a large cloud. The water vapor within the cloud as well as the water vapor from evaporating supercooled liquid droplets provides a continuous source of moisture for ice crystals, allowing them to grow rapidly. Eventually, these ice crystals become large enough to fall. The same issues exist with updrafts and rising air, but at some point the crystals will fall faster than the updrafts within the cloud.

Sometimes, ice crystals collide with nearby supercooled droplets in the cloud, causing them to freeze onto the crystal as ice. The ice crystal will grow larger and larger as it collides into more droplets, this is called **riming or accretion**. This forms an icy clump called **graupel** (also known as snow pellets), which may break apart into tiny ice particles as it collides with more droplets. These splinters may form graupel of their own as they collide into other droplets, which in turn may also splinter, causing a chain reaction.

In clouds that are colder, ice crystals may collide together and break apart into smaller ice particles, which act as tiny seeds that can freeze supercooled droplets on contact. This could also cause a chain reaction that produces many ice crystals. As these ice crystals fall, they can collide and stick together. This process of collision and sticking is called **aggregation**. A fully grown ice crystal is what we call a **snowflake**.



The growth of ice crystals occurs during its descent through aggregation ([CC 0](#)).

There must be many times more water droplets in a cloud than ice crystals for ice crystals to get large enough to fall as snow—on the order of 100,000:1 to 1,000,000:1.

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7.3: Precipitation Types

The ice phase process accounts for most global precipitation. The term **precipitation** refers to all types of precipitation—from fog, rain, snow, hail, etc. Anything composed of water that is falling in the atmosphere can be called precipitation. However, we know that not everywhere in the world gets snow all the time. While the ice phase process helps with precipitation formation, many things can happen to falling drops along their journey from the cloud to the ground. Here are a few examples.

Rain: Ice crystals melt before they hit the ground.

Snow: Ice crystals collide and stick, forming a fully grown ice crystal, and falling to the ground as a snowflake.

Graupel: Ice crystals collide and stick with other ice crystals forming clumps of snow called graupel.

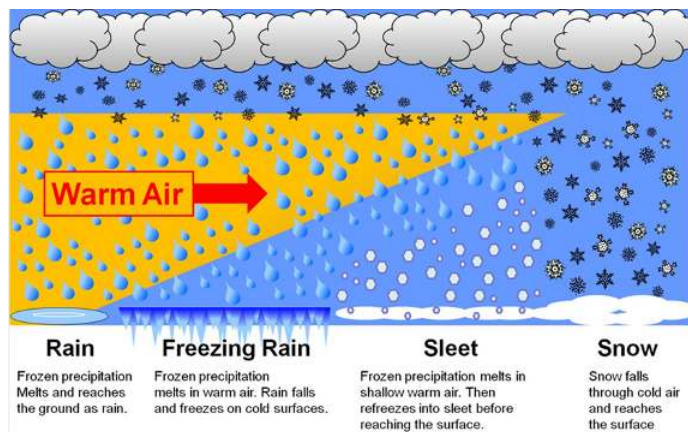
Sleet: A mixture of rain and snow, formed by partial melting.

Freezing Rain: Supercooled liquid rain that freezes on impact with the surface.

Ice Pellets: Ice crystals melt before they hit the ground, refreeze in a cold layer, usually just above the surface, and end up falling as frozen rain drops.

Hail: Ice crystals that repeatedly pass through a supercooled liquid region of cloud where riming builds up on the hydrometeor. The formation of hail requires strong updrafts and a relatively long time inside of a cloud.

As you can see, the journey of an ice crystal after formation is not always straightforward and depends strongly on the environmental conditions, especially temperature. In the following chapters (especially chapter 12), we will learn how airmasses and fronts combine together regularly in Earth's atmosphere to create conditions that are conducive to all types of precipitation.



Precipitation changes associated with the passage of a warm front ([Public Domain](#)).

The above image provides an example of how a warm front creates temperature gradients in the atmosphere that produces precipitation types from rain, freezing rain, sleet, and snow depending on your location with respect to the front.

Chapter 7: Questions to Consider

1. Which rain formation process produces most of the world's precipitation? What type of cloud does this process occur in?
2. What factors determine how large a cloud droplet can grow through collision-coalescence?
3. Match the ice nucleation methods:
- 4.

Selected Practice Question Answers:

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CHAPTER OVERVIEW

8: Weather Reports and Map Analysis

Alison Nugent and David DeCou

Learning Objectives

By the end of this chapter, you should be able to:

- Describe a few global atmospheric observation networks
- Define the meaning of pressure and temperature lines on a map
- Discuss the importance of synoptic maps, and the information required to make one

[8.1: Introduction](#)

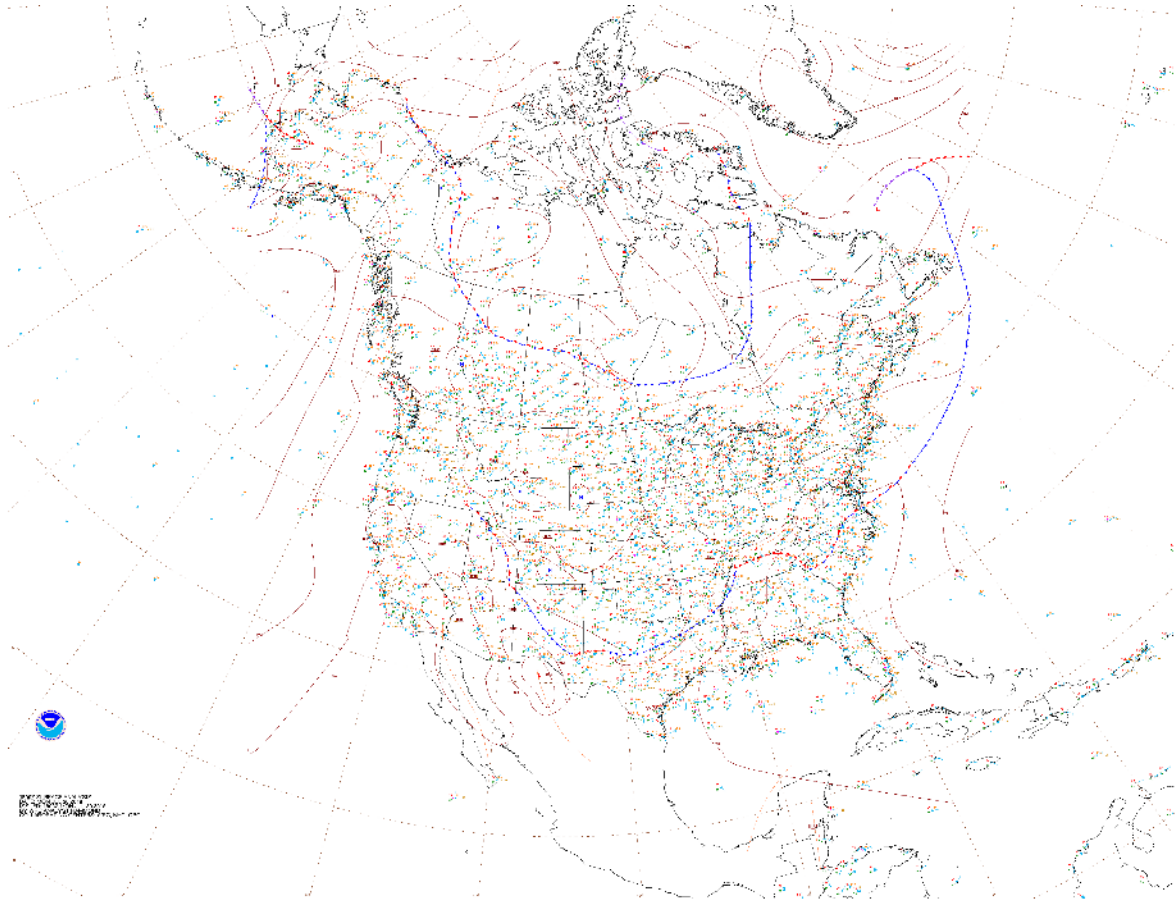
[8.2: Meteorological Reports and Observations](#)

[8.3: Synoptic Weather Maps](#)

[8.4: Map Analysis](#)

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8.1: Introduction



High resolution surface map ([Public Domain](#)).

Weather encompasses the state of the atmosphere at any given time, and it is in a constant state of change. Temperature, pressure, humidity, wind, precipitation, cloud cover, visibility—this type of data is constantly being collected around the globe, both at the surface and aloft in the upper atmosphere. How is this data collected and where does it come from? How is it shown on a map? What is the purpose of plotting weather conditions on a map if they are changing all the time? This chapter serves as a brief introduction to surface synoptic weather maps and interpreting the data plotted on them. Simplified weather charts are used frequently on TV, showing locations of high and low pressure systems, fronts, and storm systems. You're likely familiar with this material already, just from your day to day experiences.

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8.2: Meteorological Reports and Observations

The **World Meteorological Organization (WMO)** is a branch of the United Nations and it sets global weather observation standards to be used by countries across the globe. Weather systems span countries and continents so weather observations have to be synchronized to get an accurate big-picture view (a **synoptic** view) of the weather at a given time. Upper air and surface observations are taken at specific times in **Coordinated Universal Time (UTC)** so that they can be coordinated simultaneously across different time zones. For example, most upper-air observations are taken at 00 and 12 UTC, but surface observations are typically taken more frequently. Synoptic weather maps put together these observations.

Observations are reported using internationally standardized codes. One of the most commonly used codes is **METAR**, which comes from the French *MÉTéorologique Aviation Régulière*, or *Meteorological Terminal Aviation Routine*. It is a summary of surface weather conditions reported at hourly or half hourly intervals depending on the station. METAR is provided by airport terminals for the purposes of aviation meteorology. In general, METAR includes reports of wind speed and direction, visibility, cloud layers at different levels of the atmosphere, surface temperature and dew point temperature, air pressure, as well as weather conditions such as precipitation or thunderstorms near the station. **SPECI** is a special non-routine form of METAR, provided when airport weather conditions change significantly. You are not required to translate or memorize METAR for the purposes of this class, but it is useful to recognize it because it is very commonly used. Translations for parts of METAR code can be easily found online.

METAR: PHNL 072153Z 05012KT 10SM FEW025 FEW050 BKN200 31/17 A3006 RMK AO2 SLP180 T03060167

Above is an example of METAR, taken from Honolulu International Airport. “PHNL” is the ICAO airport code for Honolulu International. The “072153Z” indicates that the report was given on the 7th of the month (August 7th) at 2153Z (Universal Time), which is 11:53 AM local time. The “05012KT” is the wind report, and indicates that winds are blowing at 12 knots from 50°, which is roughly from the northeast. The “10SM” indicates that visibility is 10 statute miles, which is another way of saying that visibility on the runway is clear. If visibility is 10 statute miles or greater, it is always reported as just “10SM”. The “FEW025 FEW050 BKN200” are reports of different cloud heights, which are important for aviation purposes. This report says that there are a layer of a “few” clouds at 2,500 and 5,000 feet, with a layer of “broken” clouds at 20,000 feet. The “31/17” is telling us that the temperature is 31°C with a dew point temperature of 17°C. The “A3006” gives the station air pressure at 30.06 in Hg (inches of Mercury, this unit is still primarily used in aviation). The “RMK” denotes the remarks section of a METAR report where additional remarks and information about the weather are provided. The “AO2” is a code that indicates that the airport observing site is automated and contains a precipitation sensor. Some sites are not automated or do not have a precipitation sensor so there are different codes to denote this. Automated reports can also have non-automated remarks added to them. The “SLP180” gives the sea level pressure as 1018.0 mb. And the “T03060167” gives a more accurate temperature and dew point temperature reading. It reports that the temperature is actually 30.6°C and the dew point temperature is actually 16.7°C.

Weather Observation Locations

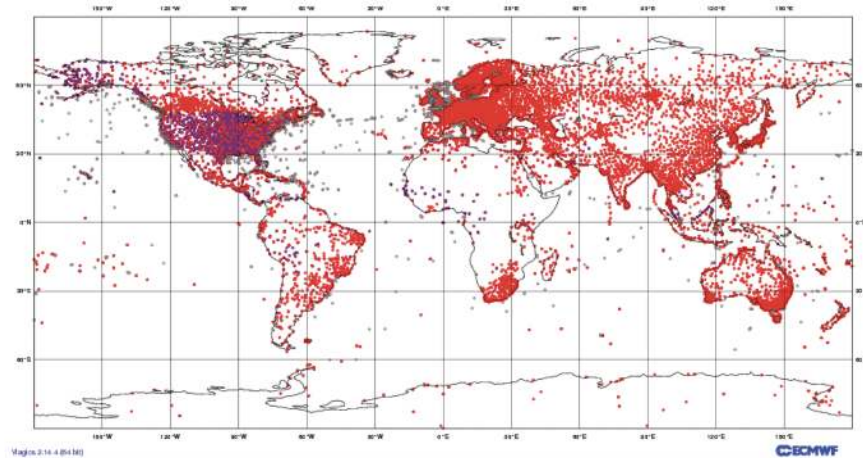
Surface weather observations include automated observations from Automated Surface Observing System (ASOS) sites in the United States, as well as hourly observations from airports around the world, reported as METAR. Manual and ship observations are also made at specific times. Surface observations of the weather are important for providing information about the weather conditions that we experience here on the ground. These conditions include temperature, dew point temperature, wind speed and direction, air pressure, cloud cover, visibility, and weather conditions such as precipitation or thunderstorms.

Although surface weather observations are important, they only tell a part of the full story. Just as ripples on the surface of a river can be a sign of what is happening below, surface observations can give an idea of what is happening above. Much of the weather that happens at or near the ground is strongly affected by conditions higher up in the atmosphere.

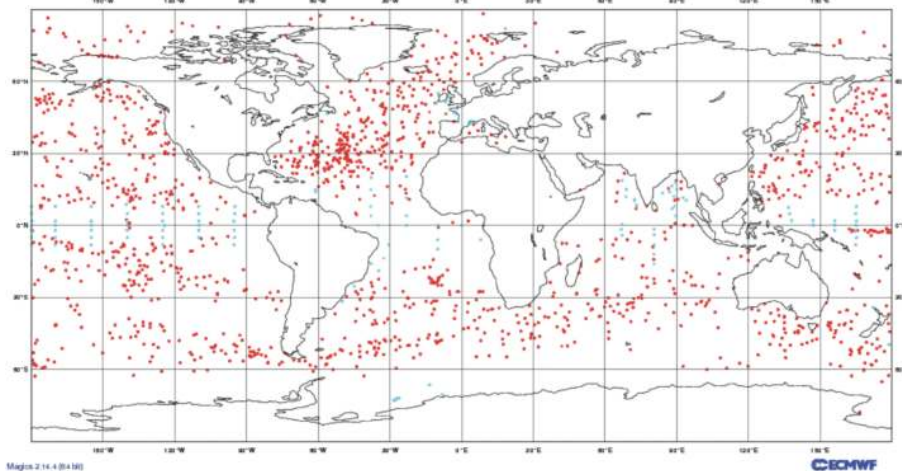
Observations of the upper atmosphere are taken by **radiosonde** packages, which are carried aloft by weather balloons. These radiosonde observations (**RAOBs**) provide **soundings** of the upper-air environment, giving information about temperature, humidity, and pressure at vertical levels throughout the atmospheric column. Recall from Chapter 5 that temperature and humidity aloft are usually plotted against pressure in a thermodynamic diagram called a Skew-T. Some radiosondes can infer wind data at different heights—these are called **rawinsondes**. When these instrument packages are dropped from an aircraft, they are called **dropsondes**. All of this data is accumulated, organized, tested, and stored in computers at governmental centers such as the

European Centre for Medium-Range Weather Forecasts (ECMWF). Other data collecting systems include weather radar (such as the **NEXRAD** network in the United States) and satellite, as well as vertical **wind profilers** and Radio Acoustic Sounding Systems (**RASS**).

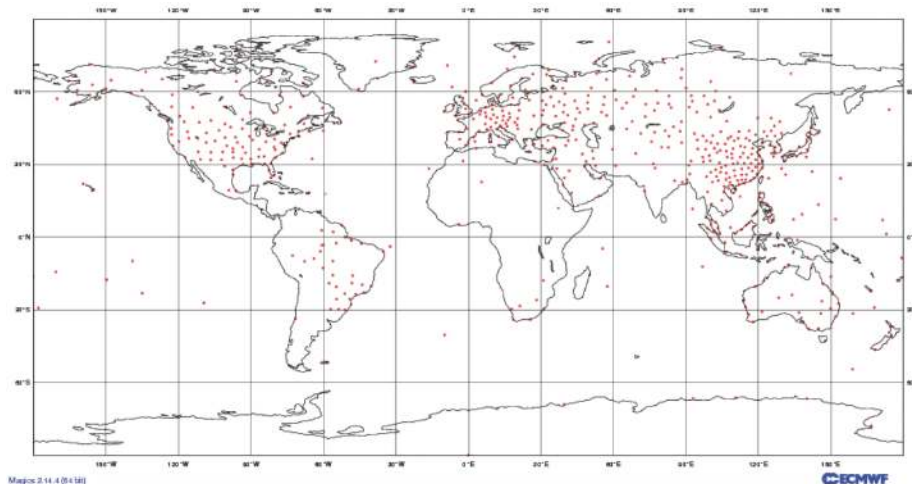
The following figures show locations of data collected by ECMWF over a 6-hour time period. The first figure shows surface observation locations located on land, the second shows surface observations over the ocean, and the third shows the upper air sounding observational network.



Surface observation locations for temperature, humidity, winds, clouds, precipitation, pressure, and visibility ([CC BY-NC-SA 4.0](https://creativecommons.org/licenses/by-nc-sa/4.0/)).



Buoy observation platforms for surface temperature and winds over the ocean ([CC BY-NC-SA 4.0](https://creativecommons.org/licenses/by-nc-sa/4.0/)).



Upper air observation and sounding locations for temperature, pressure, and humidity aloft. Stull Figure 9.4 ([CC BY-NC-SA 4.0](https://creativecommons.org/licenses/by-nc-sa/4.0/)).

Sea-Level Pressure Adjustment

Weather stations exist at many altitudes. Because air pressure decreases with height, weather stations in cities at high elevations will report far lower air pressures than cities at lower elevations. Air pressure varies vertically far more than it does horizontally. To make air pressure comparable, the station pressure needs to be adjusted to the pressure it would have if the station were at sea level. If this were not done, pressure differences between nearby surface stations would be dominated by their difference in elevation, and a surface pressure map would more closely resemble a map of elevation rather than a map of atmospheric pressure. By correcting for altitude differences, synoptic maps show mean sea level pressure, which is used to show low and high pressure centers near the surface.

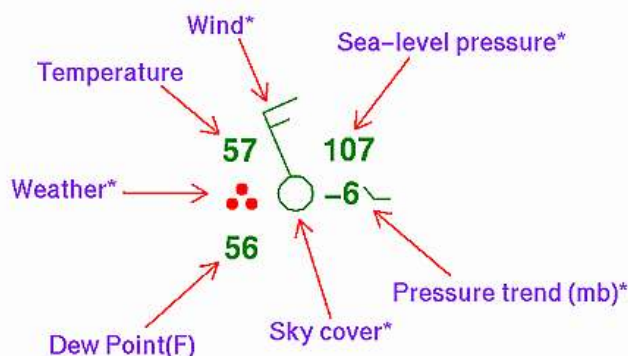
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8.3: Synoptic Weather Maps

Surface weather observations taken at many weather stations at a given time can be shown on a weather map through the use of a **station plot model**, an example of which is given below. Station plot models typically provide wind information through wind barbs, as well as current weather conditions, sea level pressure, temperature, dew point temperature, visibility, and cloud cover. Sometimes additional information such as pressure tendency, cloud type, and precipitation amounts are also given.

When reading wind barbs, keep in mind that the flag always points in the direction that the wind is coming from. You can think of it as an arrow flying from a bow, the flag represents the feathers, which are at the back of the arrow in the direction the arrow is coming from. The tip is the direction the arrow is flying toward. More information on wind barb and cloud cover symbols are provided below.

Sea level pressure is typically given in three digits, with the last digit being the nearest tenth. The initial 9 or 10 is left out. So, “147” is actually 1014.7 mb and “998” would be 999.8 mb. When reading station plot pressure, mentally place a decimal point to the left of the third digit, and use your intuition and the pressure of nearby stations in order to decide if the pressure has a leading 9 or 10.



Example station plot model ([Public Domain](#)).

The table below provides information for wind barbs. Two concentric circles with no lines indicates calm wind and a line with no barbs is 1-2 speed units. A half line is 5, a full line is 10, and a triangle is 50. On many weather maps, the unit for winds is given in knots, although different weather maps can use different units so it is always good to check. The lines and triangles are read a bit like roman numerals—just add them up!

Symbol	Wind Speed	Description
	calm	two concentric circles
	1 - 2 speed units	shaft with no barbs
	5 speed units	a half barb (half line)
	10 speed units	each full barb (full line)
	50 speed units	each pennant (triangle)
• The total speed is the sum of all barbs and pennants. For example,  indicates a wind from the west at speed 75 units. Arrow tip is at the observation location. • CAUTION: Different organizations use different speed units, such as knots, $m\ s^{-1}$, miles h^{-1}, $km\ h^{-1}$, etc. Look for a legend to explain the units. When in doubt, <u>assume knots</u> — the WMO standard. To good approximation, $10\ knots \approx 5\ m\ s^{-1}$.		

Wind barb chart ([CC BY-NC-SA 4.0](#)).

The table below provides sky cover information, which is typically shown in the circular center of the station plot model. Depending on the symbol type, it can indicate clear sky or various levels of overcast skies.

Sky Cover (oktas)	Sym-bol	Name	Abbr.	Sky Cover (tenths)
0		Sky Clear	SKC	0
1		Few* Clouds	FEW*	1
2				2 to 3
3		Scattered	SCT	4
4				5
5		Broken	BKN	6
6				7 to 8
7				9
8		Overcast	OVC	10
(9)		Sky Obscured		un-known
(/)		Not Measured		un-known

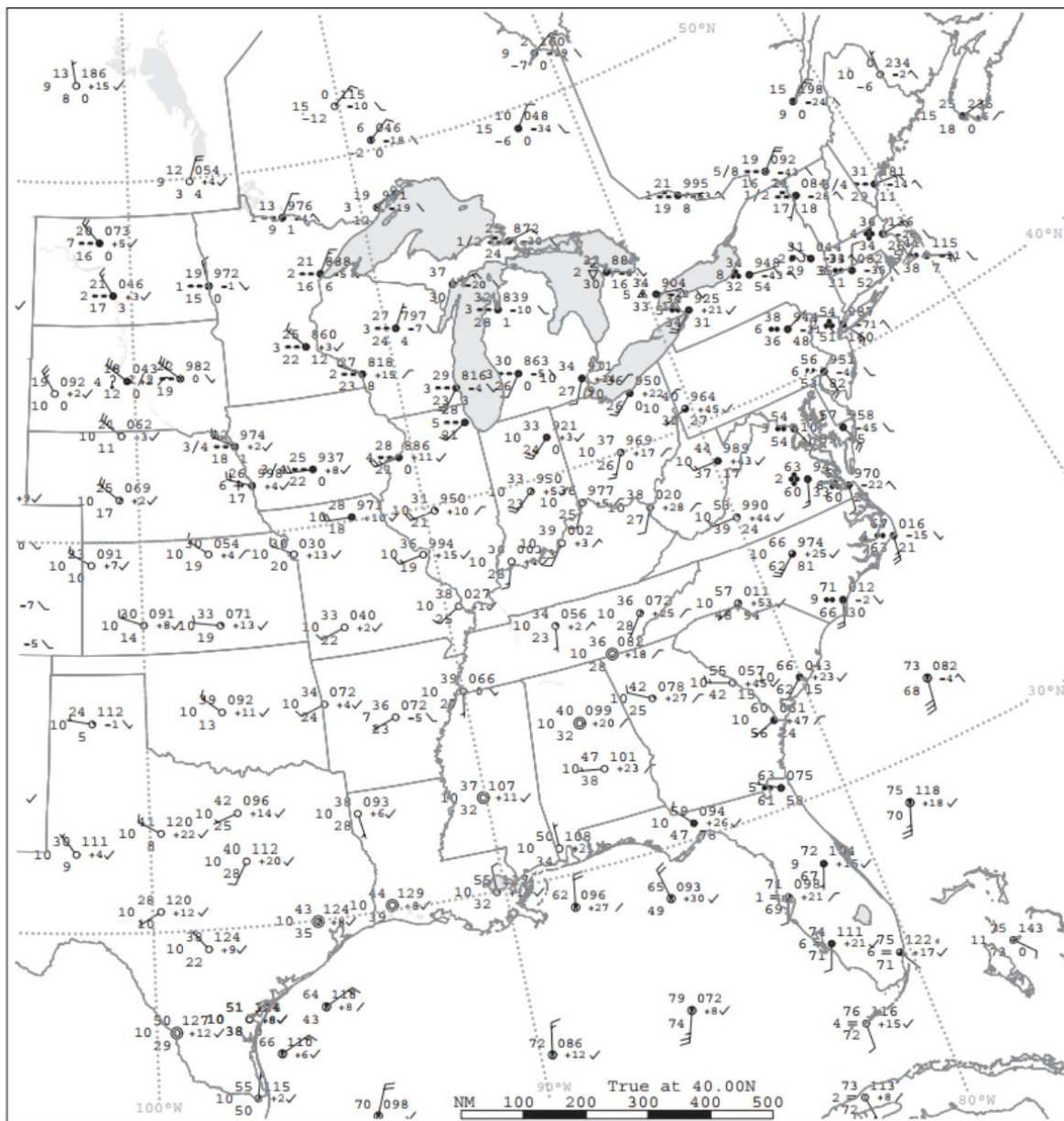
* "Few" is used for $(0 \text{ oktas}) < \text{coverage} \leq (2 \text{ oktas})$.

Sky cover chart. Stull Table 9-10 ([CC BY-NC-SA 4.0](#)).

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8.4: Map Analysis

Surface observations as described above are plotted on a map in the form of station plot models, as in the figure below. When looking at the map of raw data below, you can get an idea of the prevailing wind direction in different parts of the US, which areas are experiencing widespread rain, which areas are overcast, and from the wind direction patterns you can even infer centers of high and low pressure. However, much of this information is not very clear without doing some analysis on the map. Map analysis can either be done by hand or by computer, and involves the drawing of **contours**, or **isopleths** (lines of equal value) to connect areas of constant air pressure and temperature at the surface. Aloft, contours may be drawn to show areas of constant height, constant humidity, constant wind speed, and other parameters of interest at constant pressure levels. Map analysis also includes drawing and labeling boundaries in the atmosphere, such as **fronts** or **dry lines** to show the locations and movement of air masses. Fronts will be covered in a later chapter. Here, we will focus on lines of constant pressure, **isobars**, and lines of constant temperature, **isotherms**.



Map with station plot models (CC BY-NC-SA 4.0).

Isopleths

The below figure shows a grid of equally spaced temperature values in °C. When analyzing the temperature on a map, you draw isopleths of constant temperature: isotherms. The main purpose of this is to separate areas of warmer air from cooler air, and to get an idea of how quickly the temperature changes across a horizontal distance, which is also known as the temperature gradient. Isotherms are typically drawn every 5°C on the 5's – for example, 5°C, 10°C, 15°C, 20°C, 25°C, and so on. Standard conventions are used for contouring on different map surfaces, for example, isotherms are usually dashed and shown in red. You will not need to know these conventions for this class, but it may still come in handy to be aware.

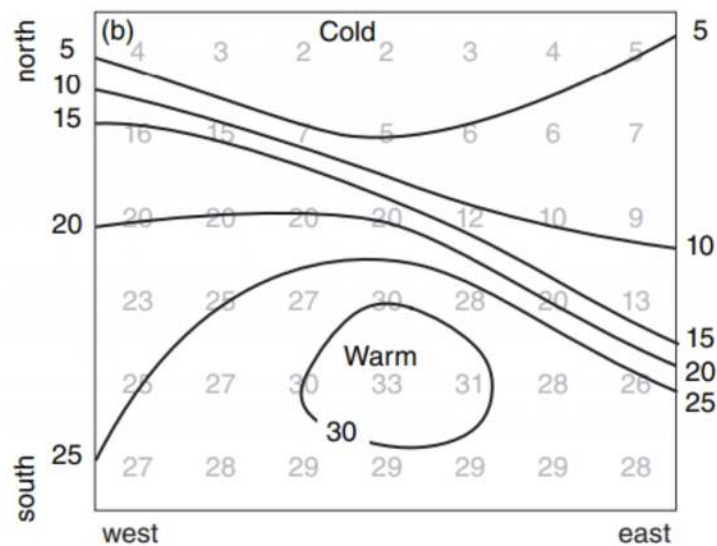
In looking at the below temperature field, you can see that there are many values that do not line up exactly with the lines you need to draw. Many values lie in between contours, such as 4°C and 16°C. You will need to interpolate data, meaning you will have to infer where a datapoint is based on the data around it. Starting from the upper left corner, imagine that you are going to draw your 5°C isotherm. Five lies much closer to 4 than 16, so you will start your isotherm close to the 4. 5°C will also lie in between 3 and 15 °C, but slightly farther away from the 3 value, so you will slope your isotherm down slightly. Five degrees centigrade will also lie in between 2 and 7 °C, so you will continue your isotherm to the right. However, it will slope down even further, because 5 lies closer to 7 than to 2. Your isotherm will then cross directly through the 5°C data point and slope back upward slightly between the 3 and 6 and directly through the middle between the 4 and 6 and back up through the 5 in the upper right hand corner.

(a)

north	4	3	2	2	3	4	5
	16	15	7	5	6	6	7
	20	20	20	20	12	10	9
	23	25	27	30	28	20	13
	25	27	30	33	31	28	26
south	27	28	29	29	29	29	28
	west						east

Blank temperature field ([CC BY-NC-SA 4.0](#)).

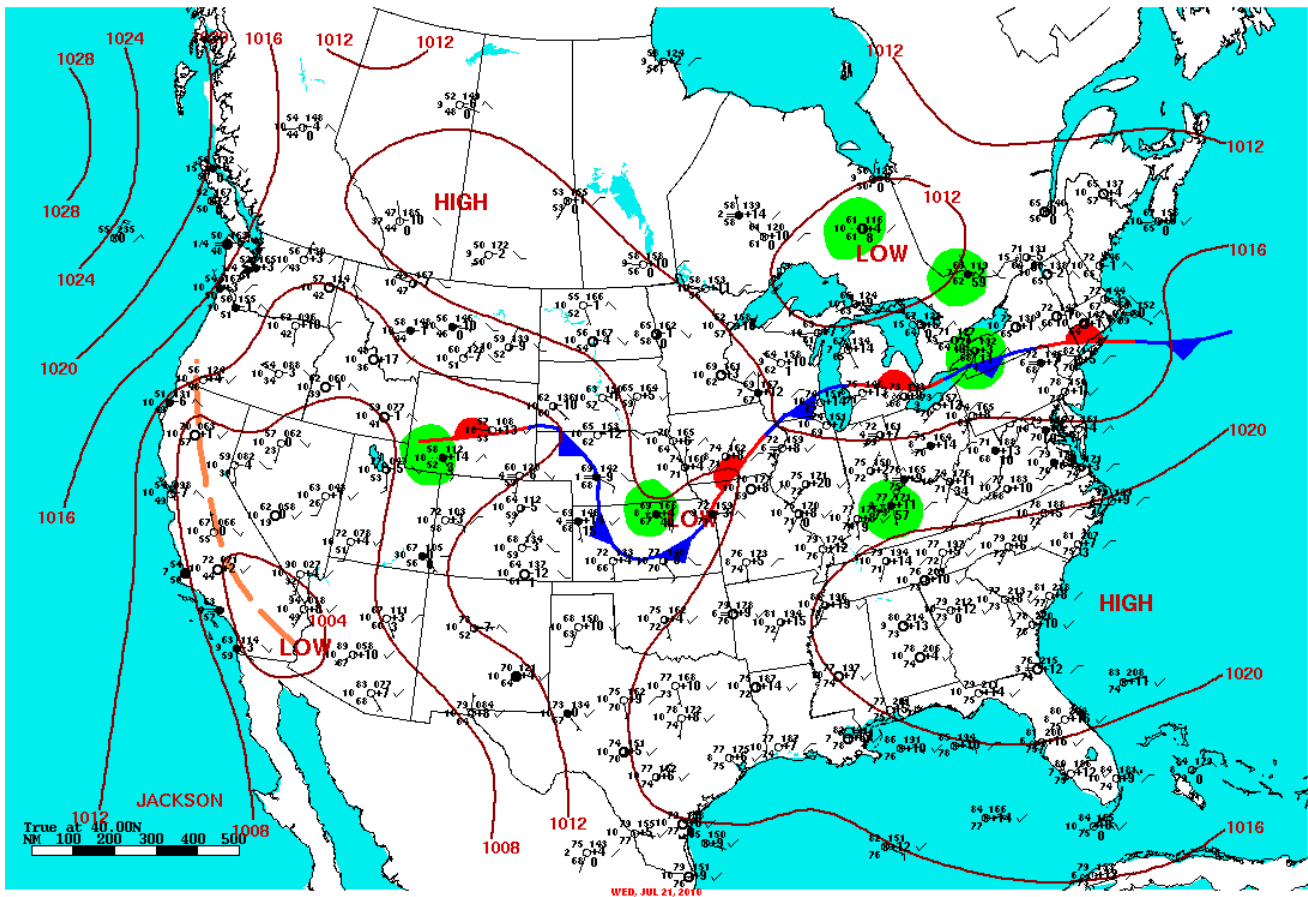
The complete contoured example of this temperature field is given below. The atmosphere is a continuous fluid, so any fields that you are contouring (pressure, temperature) will never have values that jump or suddenly end. Contours will either be closed (both ends will connect) or they will extend to the edge of the map. They will *never* cross or end suddenly. If two contours were to cross, it would mean that one place has two different air temperatures at one time, which is impossible. Areas where isotherms are close together indicate a strong temperature gradient, which may be indicative of a frontal zone.



Completed temperature field with isotherms ([CC BY-NC-SA 4.0](https://creativecommons.org/licenses/by-nc-sa/4.0/)).

Isobars

Pressure fields are analyzed by drawing **isobars**, or lines of constant air pressure. The surface mean sea level pressure is analyzed in much the same way as the isotherms above. However, the conventions are different. Typically, pressure is analyzed every 4 mb and centered on 1000 mb. Isobars are typically drawn as solid black lines. The mean sea level pressure field is analyzed in order to identify areas of high and low pressure, as shown in the figure below. Areas where isobars are drawn close together indicate places where the pressure gradient force is strong. A stronger pressure gradient force is indicative of stronger wind speeds. As you will learn later, winds tend to flow counterclockwise around areas of lower pressure, and clockwise around areas of high pressure in the Northern Hemisphere.



Surface Weather Map and Station Weather at 7:00 A.M. E.S.T.

Surface map with isobars and station plot models ([Public Domain](#)).

Because weather information is shared across continents and across country borders, it is important to maintain standards such that everyone effectively speaks the same language. Weather is one of the greatest international collaborations ever. If it weren't for China and Russia sharing their weather information with the US, our long range forecasts would be more inaccurate because we wouldn't know the initial conditions of the atmosphere upstream of the North American continent.

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CHAPTER OVERVIEW

9: Atmospheric Forces and Wind

Alison Nugent and Shintaro Russell

Learning Objectives

By the end of this chapter, you should be able to:

- Define wind and why it occurs
- Use u , v , and w to describe motion
- Describe the five physical forces that can act on a parcel of air
- Draw force diagrams for geostrophic wind, gradient wind, and wind in the atmospheric boundary layer (ABL)
- Compute the speed of a geostrophic wind, and be able to qualitatively estimate the wind speed from isobars

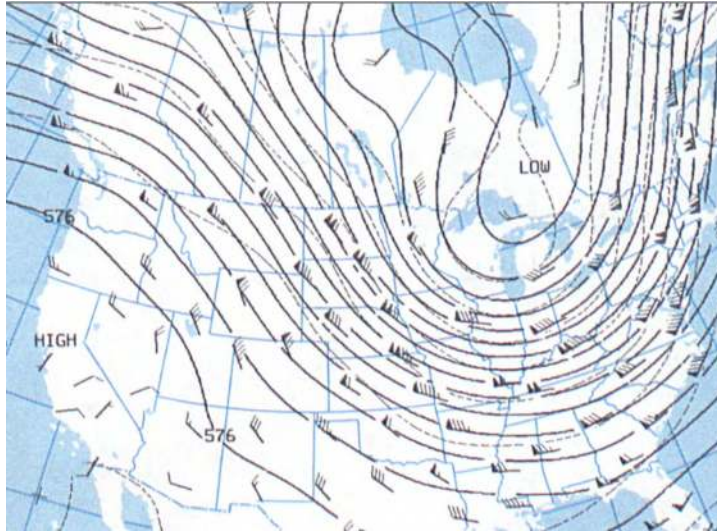
[9.1: Introduction](#)

[9.2: Main Forces](#)

[9.3: Force Balances](#)

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9.1: Introduction



Map of wind barbs and isobars ([Public Domain](#)).

Wind is the movement of air relative to the Earth's surface. As with all moving things, it is caused by a force acting on it. Force is a pull or push that changes the resting state, motion, or direction of an object. Force can also cause objects to accelerate. Human skin can sense wind when an uncountable number of molecules collide with us as they flow along in the air, and sense the pressure changes in the air flow.

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9.2: Main Forces

There are **five forces** that influence the speed or direction of horizontal winds.

1. Pressure gradient force
2. Advection
3. Centrifugal force
4. Coriolis force
5. Turbulent drag

Remember from Chapter 1 that according to the Cartesian coordinates, x points **east**, y points **north**, and z points **upwards**. To define wind, we use wind components u , v , and w which correspond to the x , y , and z directions. These wind components are used in equations of motion used to predict wind to designate direction in a three-dimensional plane. We'll go through each of the five forces one by one to discuss how they affect wind speed and/or direction.

Pressure Gradient Force

A **pressure gradient** (PG) is a change in pressure over a distance. Therefore, the units of the pressure gradient force are Pascals per meter ($P\ m^{-1}$). The pressure gradient can be calculated simply as the change in pressure divided by the distance over which that change occurs. The size or strength of the pressure gradient determines the size or strength of the force that results from it.

$$PG = \frac{\Delta P}{\Delta x}$$

The **pressure gradient force** (PGF) is a force from high to low pressure over a distance. Without differences in pressure, there would be no wind because there would be nothing to accelerate airflow. In Chapter 9 we learned about isobars, or lines of constant pressure. When isobars are tightly packed, we know that there is a strong pressure gradient or large change in pressure over a relatively short distance. A strong pressure gradient results in a large pressure gradient force and, as we'll see, higher wind speeds. Depending on the text you read, you'll see a number of different forms of the pressure gradient force. Two equivalent representations are given below.

$$\frac{PGF_x}{m} = -\frac{1}{\rho} \cdot \frac{\Delta P}{\Delta x}$$

The PGF as defined above is simply a change in pressure divided by the distance and air density. This gives the force per unit mass, hence PGF/m . Above we define PGF in the x -direction, but it can act in any horizontal plane. The units of the PGF will be in units of acceleration above because Force = Mass * Acceleration, but because we're dividing the left-hand side by Mass the units will be $m\ s^{-2}$. The negative signs in the above equation is due to the fact that the PGF acts from high pressure to low pressure.

Another equivalent way to represent the PGF is as follows.

$$|PGF| = \text{Volume} \cdot \frac{\Delta P}{L}$$

Because $Mass = Volume * Density$, we then see that $Volume = Mass/Density$. The second equation is identical to the first, except that $Mass/Density$ is on the right hand side in the form of $Volume$ instead of being split between left and right sides of the equals sign. Also, the variable L is used for length or distance instead of Δx so that it isn't for a specific direction. Similarly, the absolute value of the PGF is calculated, and the direction is determined, by the location of High and Low pressure systems. Typically in class we won't calculate the PGF , only the PG . However, it'll be widely used in force balance equations.

Advection

While **advection** is included in the list of forces, it is actually not a true force. Still, advection can result in a change of wind speed in some locations. Wind moving through a point carries specific momentum, which is defined as momentum per unit mass. Recall that momentum is mass times velocity. Specific momentum then is simply equal to the velocity or wind speed. Therefore, as wind moves by a point, the wind can move or advect variations in winds to the fixed location.

Centrifugal Force

The **centrifugal force** is an apparent force that includes the effects of inertia for winds moving along a curved path. The directionality of the centrifugal force points *outward* from the center of the curve. The centrifugal force is the opposite of the centripetal force. As we know, inertia is the physical tendency to remain unchanged. Therefore inertia causes an air parcel to “want” to move along a straight line. Turning the air parcel along a curved path requires a centripetal force that pulls inward to the center of rotation. As a result, a net imbalance of other forces occurs.

You have felt the centrifugal force many times in your life. The centrifugal force is easily felt as you travel in a moving vehicle around a corner. The force that you feel pulling you outwards is the centrifugal force.

Coriolis Force

The **Coriolis force** (CF) is another apparent force that occurs due to the rotation of Earth. The Coriolis force is a deflecting force. It acts only on objects already in motion. Therefore it cannot create wind, but it can change the wind direction by deflecting it. The Coriolis force acts perpendicular to the direction of motion, but whether the Coriolis force acts 90° to the right or left of the motion vector depends on the hemisphere on Earth. In the Northern Hemisphere, the Coriolis force acts 90° to the right of the motion vector while in the Southern Hemisphere, the force acts 90° to the left of the motion vector. The equation below gives the Coriolis force

$$CF = m \cdot 2 \cdot u \cdot \Omega \cdot \sin(\phi)$$

where m is for the mass of the object in kg, u is the speed of the object in m s^{-1} , the symbol ϕ denotes latitude in degrees, and the angular rotation rate, Ω , is found from the rotation rate of Earth. Earth turns 2π radians over 24 hrs, so $2\pi/24$ hrs gives $\Omega = 7.27\text{E-}5 \text{ radians}\cdot\text{s}^{-1}$. Based on this, we can see that the Coriolis parameter will be 0 at the equator when $\sin(0) = 0$ and maximized at the poles when $\sin(90) = 1$.

We can also see that the Coriolis force is strongly dependent on the speed of the object. If we assume the “object” is actually wind, stronger winds will be more strongly deflected by the Coriolis force.

Turbulent Drag

Turbulent drag occurs when Earth’s surface or objects on it cause resistance to airflow and reduce the wind speed. Any object on Earth’s surface can cause drag, such as grass, trees, and buildings, which block and decelerate wind. The bottom layer of the troposphere around 0.3 to 3 km thick is called the **atmospheric boundary layer (ABL)**. Turbulence in the ABL mixes the extremely slow movement of air near the surface with the faster movement of air in the ABL and slows the wind speed in the entire ABL.

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9.3: Force Balances

The five forces from above affect aspects of horizontal wind speed and direction, and result in a number of common force balances found throughout Earth's atmosphere.

Geostrophic Balance

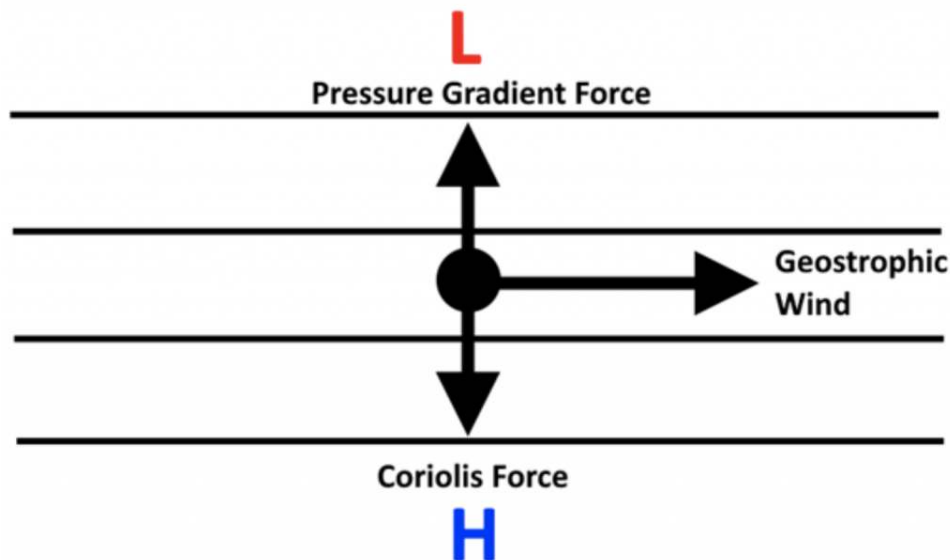
Geostrophic balance is arguably the most important force balance in the atmosphere and holds nearly all the time, except for a few specific cases scenarios to be discussed later. When in geostrophic balance, wind in the atmosphere has a balance between the pressure gradient force and the Coriolis force. In **geostrophic balance**, $PGF = CF$. The resulting wind is called a geostrophic wind. Setting the equation for CF and PGF equal to each other and solving for u gives the following equation for U_{geos} .

$$U_{geos} = \frac{PG}{2 \cdot \rho \cdot \Omega \cdot \sin(\phi)}$$

Because geostrophic winds are dependent on the pressure gradient, geostrophic winds are faster when isobars are closely spaced.

A number of assumptions are implicit to geostrophic balance. Geostrophic balance applies only under the following conditions: large temporal (>12 hrs) and large spatial (> a few km) scales; above the ABL when no surface friction is acting on the air; winds are steadily moving in a straight direction (no acceleration, negligible vertical velocity); finally, because the Coriolis force is important for the balance, it cannot hold at the equator when the CF is 0. The typical bounds are often given as >2° latitude.

The path of the geostrophic wind is parallel to the isobars. In the Northern Hemisphere, the wind direction is parallel to the straight isobars with the low pressure to the left side of wind. In the Southern Hemisphere, the direction is parallel to the straight isobars with the low pressure to the wind's right. The image below shows the force balance present in a geostrophic wind in the northern hemisphere.



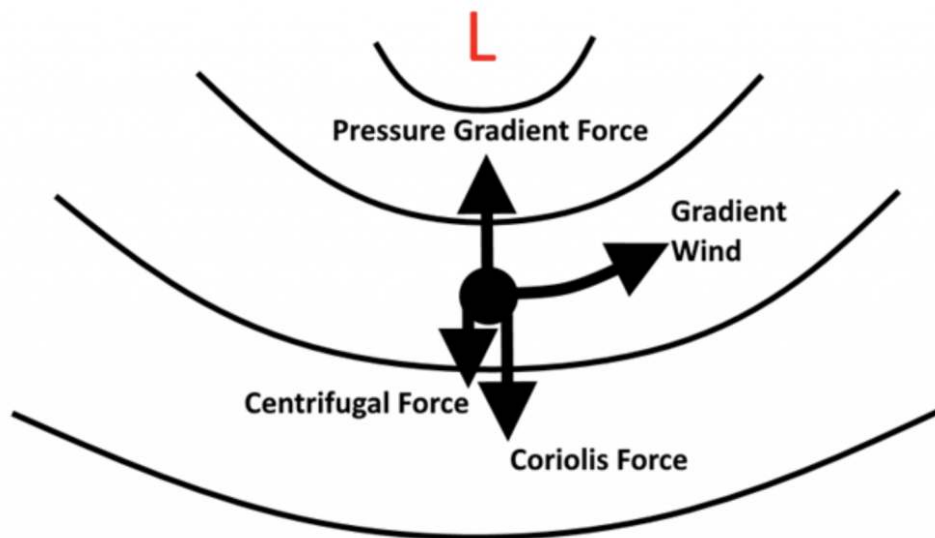
Geostrophic wind force diagram in the northern hemisphere (Image Created by Shintaro Russell via Paint.net).

To get into geostrophic balance, moving air will undergo **geostrophic adjustment**. First, air feels the pressure field (PGF) and begins moving from high to low pressure. Next, the Coriolis force (CF) deflects the object's direction once it is in motion. Finally, the air finds itself in a balance between the PGF and the CF moving parallel to the isobars instead of across them.

Gradient Wind

This next force balance applies when air is not moving in a straight line. **Gradient winds** are winds flowing along curved isobars. Winds typically blow along isobars, even if they are curved, but a different name is needed because the force balance includes one more component. Compared to geostrophic winds, gradient winds feature a balance between the Coriolis force, the pressure

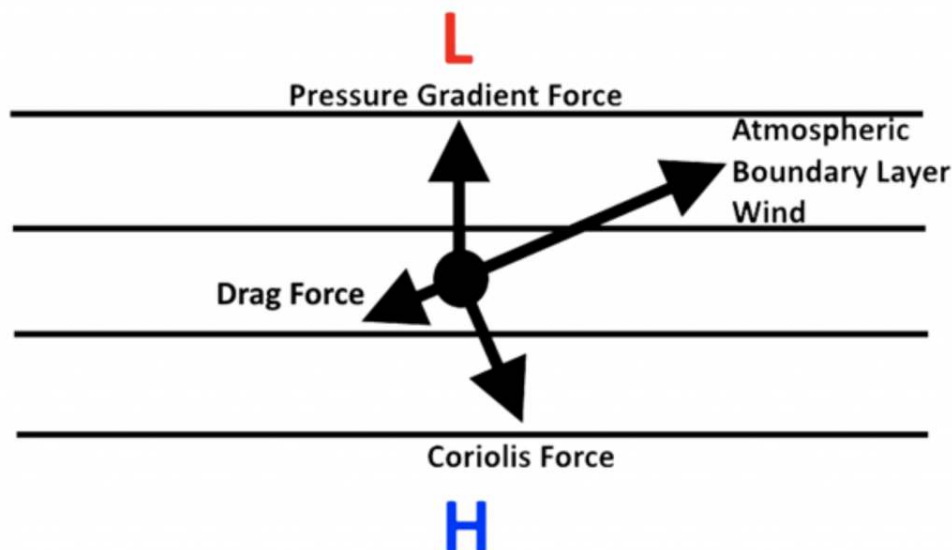
gradient force, and the centrifugal force. The centrifugal force arises because the air is flowing on a curved path. The centrifugal force acts in the same direction as the coriolis force, opposite the pressure gradient force.



Gradient wind force diagram (Image Created by Shintaro Russell via Paint.net).

Atmospheric Boundary Layer

Balanced wind in the atmospheric boundary layer (ABL) occurs when there is a balance between the pressure gradient force, Coriolis force, and the frictional drag force. Both wind shear turbulence and convective turbulence cause drag, which results in the ABL wind being slower than geostrophic (subgeostrophic), and causes the wind to cross isobars toward the low pressure.



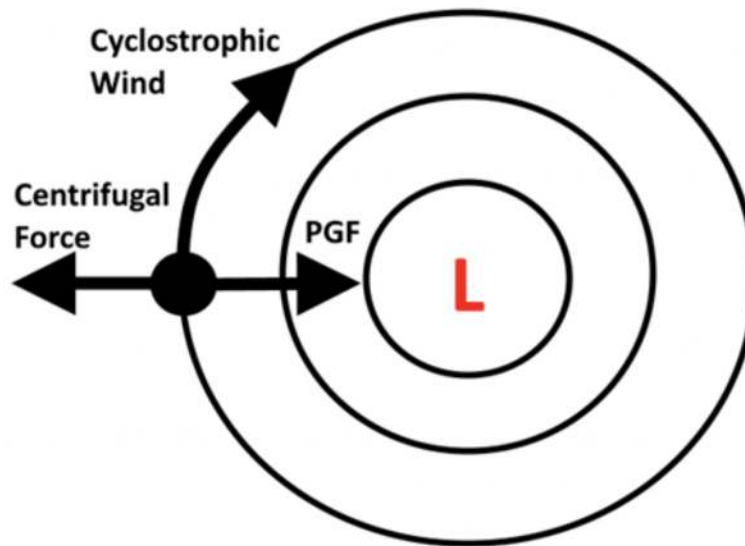
Atmospheric boundary layer wind force diagram (Image created by Shintaro Russell via Paint.net).

Again, the frictional drag force acts in the plane of motion and slows down the wind speed. The pressure gradient force doesn't change, but because the wind speed is slower, the Coriolis force is weaker. When that happens the wind cannot balance the pressure gradient force, it is pulled more by the pressure gradient force, and turns toward the low pressure.

Cyclostrophic Wind

Cyclostrophic wind occurs at smaller cyclonic scales (at the mesoscale) such as tornadoes, waterspouts, and even the center of a tropical cyclone. Because the scale is small, the Coriolis force does not play a role. When a small cyclonic scale such as a tornado first forms, both tangential winds and centrifugal force increase much faster than the Coriolis force due to the very strong pressure

gradient force. As a result, centrifugal force balances with the pressure gradient force, ignoring the negligible effects of Coriolis force. Because the scale is small and independent of the Coriolis force, the direction of cyclostrophic winds can be either clockwise or counterclockwise in both hemispheres. For anticyclones or highs, however, they do not typically have strong pressure gradients. Thus, winds around the high are too weak to be in cyclostrophic balance.



Cyclostrophic wind force diagram where the pressure gradient force is balanced by the centrifugal force (Image Created by Shintaro Russell via Paint.net).

All of the wind balances discussed (geostrophic balance, gradient wind, ABL wind, and cyclostrophic wind) occur in Earth's atmosphere under differing conditions. The following chapters will make these applications clearer and you can check back here for reference.

Chapter 10: Questions to Consider

1. Explain why wind occurs.
2. What are u , v , and w ?
3. Which forces influence the direction and speed of horizontal winds?
- 4.
5. Drag the terms to their correct position:

Selected Practice Question Answers:

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CHAPTER OVERVIEW

10: General Circulation

Alison Nugent and David DeCou

Learning Objectives

By the end of this chapter, you should be able to:

- Describe the differential heating Earth experiences, and how heat is redistributed
- Diagram vertical atmospheric circulations (Hadley cell, Ferrel cell, Polar cell)
- Diagram surface wind directions (trade winds, belt of westerlies, etc.)
- Discuss the distribution of heat over Earth's surface and how it drives global circulation, including its connection to the Coriolis force

[10.1: Introduction](#)

[10.2: Differential Heating](#)

[10.3: Single-Cell Model](#)

[10.4: Three-Cell Model](#)

[10.5: Global Surface Winds](#)

[10.6: Jet streams](#)

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10.1: Introduction

The focus of this chapter is on the typical wind circulations found on Earth as a result of the forces affecting wind in the atmosphere, as introduced in Chapter 10. The average global winds are called the **general circulation of the atmosphere**. To find typical wind circulations, one needs to average wind speed and duration over a long period of time. Averaging over time removes short duration fluctuations, allowing the primary sense of movement to be visualized.

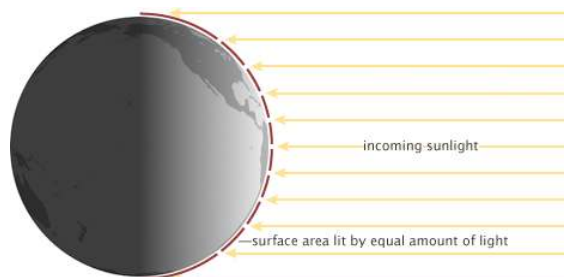
The reason we have global wind patterns is ultimately due to a differentially heated, rotating Earth. The differential heating of Earth continually causes an imbalance in air pressure and temperature around the world, which in turn causes a continuous general circulation of winds that attempt to restore balance.

While actual winds in a given place and time may differ from the average general circulation, the average can provide an explanation for how and why the winds prevail from a particular direction in a certain place. For example, the prevailing surface winds tend to be westerly in the continental United States, but northeasterly in Hawai'i. The general circulation also serves as a model for how heat and momentum are transported from the equator to the poles.

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10.2: Differential Heating

Because the Earth is round, solar radiation is not equally spread at all latitudes. Near the equator where sunlight shines directly on Earth, more solar radiation per square meter is received as compared to near the poles where sunlight shines at sharp angles to the surface (see image below). Toward Earth's poles, the same solar radiation is spread over a larger surface area such that each square meter of Earth's surface gets less radiation at the poles. As Earth rotates, the incoming solar radiation is zonally spread along latitude lines.

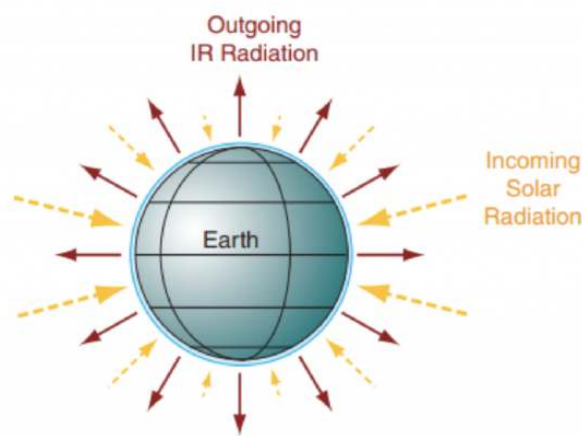


Earth's uneven heating by the sun due to the curvature of its surface [NASA \(Public Domain\)](#).

In this way, incoming solar radiation depends on latitude. The sun shines more directly on tropical regions at lower latitudes than at higher latitudes all year-round. Solar radiation adds heat to the Earth-atmosphere-ocean system, and thus lower latitudes get heated more than higher latitudes. This should be as expected because we know the tropics are warmer than the polar regions.

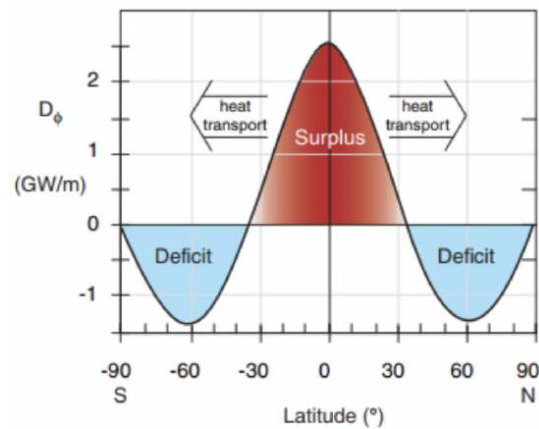
While Earth is continually heated by the sun, it is also continually losing energy by emitting outgoing longwave infrared (IR) radiation at all latitudes, and at all times, both on the light and the dark side of the globe. You'll recall from Chapter 2 that IR radiation is emitted by Earth according to the Stefan-Boltzman law, which is highly dependent on temperature.

When averaged over the globe and over long time scales, incoming UV radiation exactly balances outgoing IR radiation. But, latitude by latitude, incoming UV and outgoing do not perfectly balance. More solar energy is received by the Earth in the tropics, and while the cooling by outgoing IR radiation helps to offset this, there is still a net gain of radiative energy in the tropics. However, near Earth's poles, incoming solar radiation is less direct and too weak to offset the cooling by outgoing IR radiation, so there is net cooling at the poles. This causes warmer air at the equator, and cold air at the poles and drives Earth's atmospheric general circulation.



Incoming solar radiation (yellow dashed) and outgoing infrared radiation (red solid) diagram ([CC BY-NC-SA 4.0](#)).

The above image illustrates this more directly. Incoming solar radiation is focused near the equator, while outgoing IR radiation is relatively evenly spread across all latitudes. This results in the below energy surplus near the equator and deficit toward the poles.



Heat transport balances the net radiative imbalances ([CC BY-NC-SA 4.0](#)).

Earth's general circulation attempts to redistribute heat around the globe and rebalance the energy imbalances inherent in an unevenly heated, rotating planet. However, the general circulation cannot instantly balance global temperature, especially when the uneven heating is continuous. Therefore, a meridional temperature gradient always remains.

In an attempt to balance out Earth's incoming and outgoing energy, warm air is transported toward the poles, while cool air flows back toward the equator. This seems simple enough. However, this seemingly simple flow is complicated by many factors, including Earth's rotation, the position of continents, interactions with the oceans and many others. In order to build an understanding of this process, we'll start with some simplified models and build complexity.

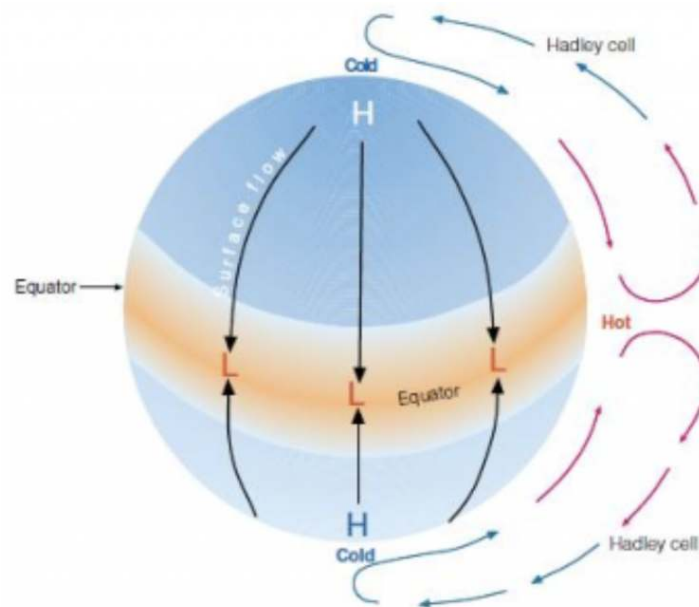
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10.3: Single-Cell Model

The first model we'll examine is the single-cell model. With this model, we make the following assumptions.

1. The earth is entirely covered with water. This is to remove any land-sea interactions.
2. There are no seasons and the sun is always shining directly over the equator. This removes seasonal wind shifts.
3. There is no Coriolis force. While the Earth rotates to spread heat along latitudinal lines, this allows us to only be concerned with the pressure gradient force.

With these assumptions in place, Earth's global circulation would like the figure below, with one giant vertically overturning cell in each hemisphere. The excess heating at the equator is transported poleward by rising warm air, which is replaced by cold sinking polar air moving equatorward. This circulation is known as the **Hadley cell**. The Hadley cell is known as a *thermally direct* circulation because in it, warm air is rising and cold air is sinking.



The single-cell model of Hadley cells on a planet ([CC BY-SA 4.0](#)).

The circulation can be thought of in two ways. In the first, hot air at the equator rises because it is warm and buoyant. It reaches the tropopause, spreading laterally north and south at high elevations. To compensate for the rising air, surface air flows toward the equator, resulting in convergence and further uplift. Continuity of this circulation results in a global circulation with rising air at the equator and sinking air at the poles.

A second way to view global circulation is that the excess heating of air at the equator creates a large area of low pressure at the surface of the planet, while excess cooling at the poles creates high pressure at the surface. This global horizontal pressure gradient causes air to flow from high to low at the surface (pole to equator), where the air subsequently rises at the equator and flows back to the poles and sinks.

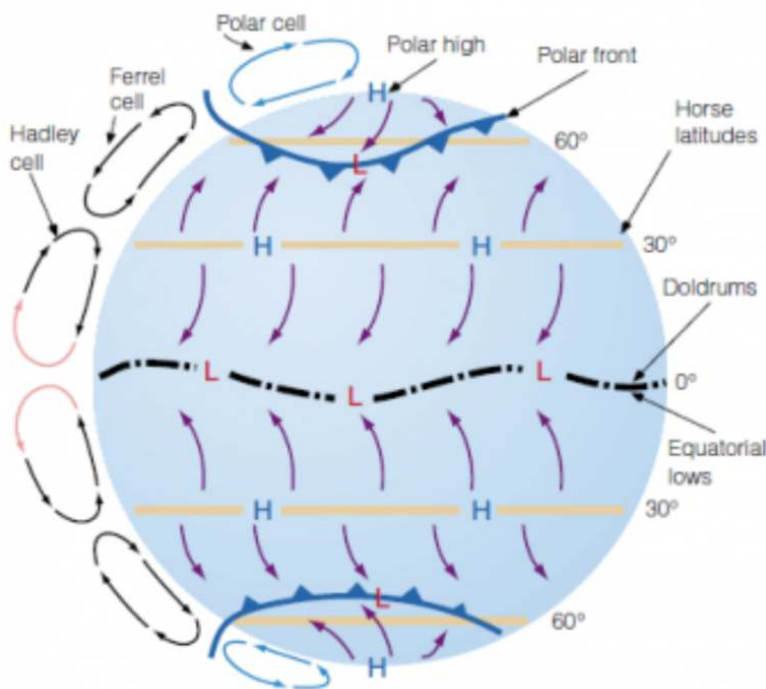
Both reasonings are plausible, it's a matter of whether you focus on temperature or pressure. The temperature differences and the resulting pressure differences are intertwined and both important for the general circulation.

While this single-cell model can explain some phenomenon and works in some ways (and on some planetary bodies), it is not the reality on Earth. Earth is a rotating planet, so we need to consider the Coriolis force in addition to the pressure gradient force. In the single-cell model, as upper level air flows from the equator toward the poles, it would be deflected by the Coriolis force. In the northern hemisphere, for example, this deflection would be toward the right resulting in a wind from west to east at upper levels. In this way, the air moving from the equator to the poles would never make it there because of the rotation of Earth. A different model is needed.

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10.4: Three-Cell Model

If we allow for the effects of a rotating planet, the simple single-cell model above breaks down into multiple cells in each hemisphere as shown in the figure below. It may look more complex and unrelated to the single-cell model, but there are many similarities from above. There is still excess heating in equatorial regions and excess cooling in polar regions. Instead of heat being redistributed by one massive Hadley cell from the equator to the poles, there are now three convective cells. The first of these is still the same thermally direct Hadley cell from before, but now it extends only from the equator to about 30° latitude. The poles still have a large high pressure system, while the equator has a large belt of low pressure along it. Let's take a closer look at what happens to the rising air just above the equator.



Three-cell model of the rotating Earth and the resulting wind circulations ([CC BY-SA 4.0](#)).

At the equator, the air near the surface is warm, winds are light, and the pressure gradient is weak. This region of monotonous weather is known as the **doldrums**. The warm air here rises, condensing into massive cumulonimbus clouds and thunderstorms, which release large amounts of latent heat as they form. The additional heat makes the air even more likely to rise, and provides the energy that drives the rising branch of the Hadley cell. This rising air reaches the stable tropopause, which blocks it from rising further, causing the air to diverge at upper levels and move poleward. Due to the Coriolis force, this upper level poleward flow is deflected to the right in the Northern Hemisphere and to the left in the Southern Hemisphere, providing westerlies aloft (near the tropopause) in both hemispheres in the Hadley cell.

As air moves poleward from equatorial regions, it is constantly experiencing radiational cooling as it emits infrared radiation. Simultaneously, this air begins to converge and pile up as it approaches the mid-latitudes (around 30° latitude in both hemispheres). This convergence of air far above the surface increases the mass of air aloft, increasing the pressure at the surface. This increase in surface pressure results in a belt of high pressure centers called **subtropical highs** around 30°N and 30°S. These latitudes are commonly known as the **horse latitudes**.

As this converging air above the subtropical highs slowly descends, it warms adiabatically by compression. This sinking air, dries the atmosphere creating generally clear skies and little rain. Over the oceans, weak pressure gradients in the high centers produce weak winds. Some of these lighter surface winds begin to move back toward the equator, and are deflected by the Coriolis force. This causes northeasterly winds in the Northern Hemisphere and southeasterly winds in the Southern Hemisphere in tropical regions. These winds are known as the **trade winds**, and they have a strong influence over the daily wind patterns in Hawai'i. Near

the equator, the northeasterly and southeasterly trade winds converge at the surface at what is known as the **intertropical convergence zone (ITCZ)**. Here, convergence further reinforces the rising branch of the Hadley cell.

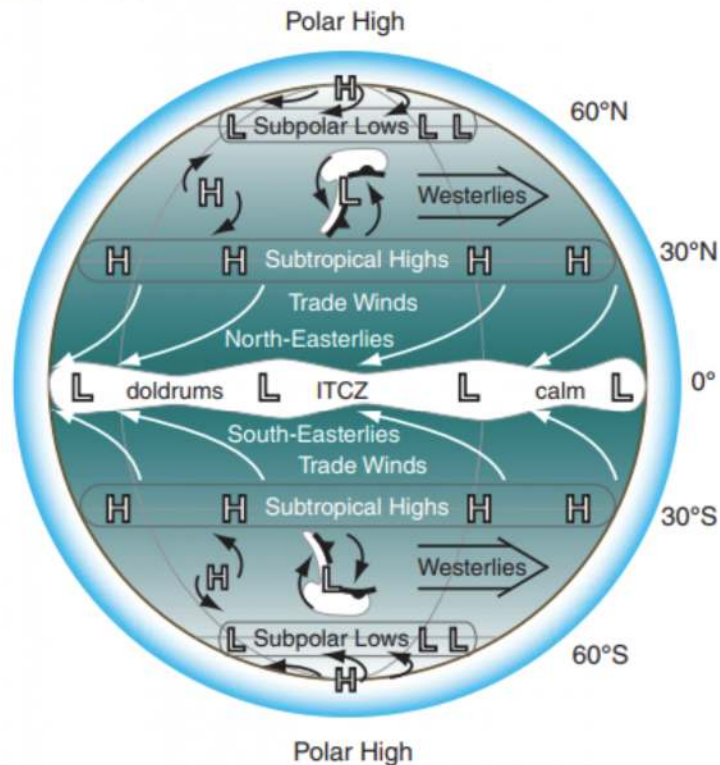
Back at 30° latitude, while some of the air sinking along the subtropical highs goes equatorward to complete the Hadley cell, some sinking air also moves poleward. This poleward moving surface air travels from from 30° to 60° and is again deflected by the Coriolis force. This results in the prevailing surface **westerlies** that impact the mid-latitudes in both hemispheres. It is for this reason that weather moves west to east across the continental US. Often, this westerly flow is interrupted by high and low pressure systems that move with the mean surface flow. We'll learn more about this in the next two chapters. As the surface air travels poleward from 30° to 60°, it collides with cold polar air moving equatorward. These air masses do not mix easily, and are separated by a boundary known as the **polar front**. At the polar front, surface air converges and rises at the **subpolar low**, and storms and convection develop here. Some of this rising air goes all the way up to the tropopause where it moves back to 30° latitude and sinks at the subtropical high along with the descending branch of the Hadley cell. This circulation cell from 30° to 60° is known as the **Ferrel cell**, which is a *thermally indirect* circulation in which cool air rises and warm air sinks.

Behind the polar front in the Northern hemisphere, cold surface polar air moves from the poles toward 60°. As the air moves equatorward, it is again deflected by the Coriolis force. In the Arctic regions, air typically flows from the northeast while in the Antarctic, air flows from the southeast. These are known as the **polar easterlies**. Along the polar front where cold polar air collides with warm air from the Ferrel cell, some of the rising air moves back toward the poles, which gets deflected as a westerly wind aloft. Eventually this air reaches the poles, sinks back to the surface, and flows back toward the polar front, which gives us the **Polar cell**.

To summarize, looking back at the three-cell model picture: there are two major belts of high pressure and two major belts of low pressure in each hemisphere (if you include the equator in both). Areas of high pressure and sinking air exist near 30° latitude and at the poles. Regions of low pressure and rising air exist over the equator and near 60° latitude by the polar front. By knowing that winds travel counterclockwise (clockwise) around low pressure systems in the Northern Hemisphere (Southern Hemisphere), and clockwise (counterclockwise) around high pressure systems in the Northern Hemisphere (Southern Hemisphere), you can get a pretty general idea of how surface winds blow around the world on average. Trade winds blow from the subtropical highs at 30° to the equator, the westerlies blow from the subtropical highs to the polar front, and the polar easterlies blow from the poles to the polar front at the surface. Areas where these winds converge will have rising motion and low pressure at the surface, and regions where these winds diverge will have sinking motion and high pressure at the surface.

How does this three-cell model match with reality? While some minor discrepancies exist, for example in reality much of the upper-level winds in the mid-latitudes are westerly like the surface, while the Ferrel cell suggests there should be easterly winds aloft. However, this model is roughly accurate for surface winds and provides a really good first order pattern for general circulation.

a) Near Surface

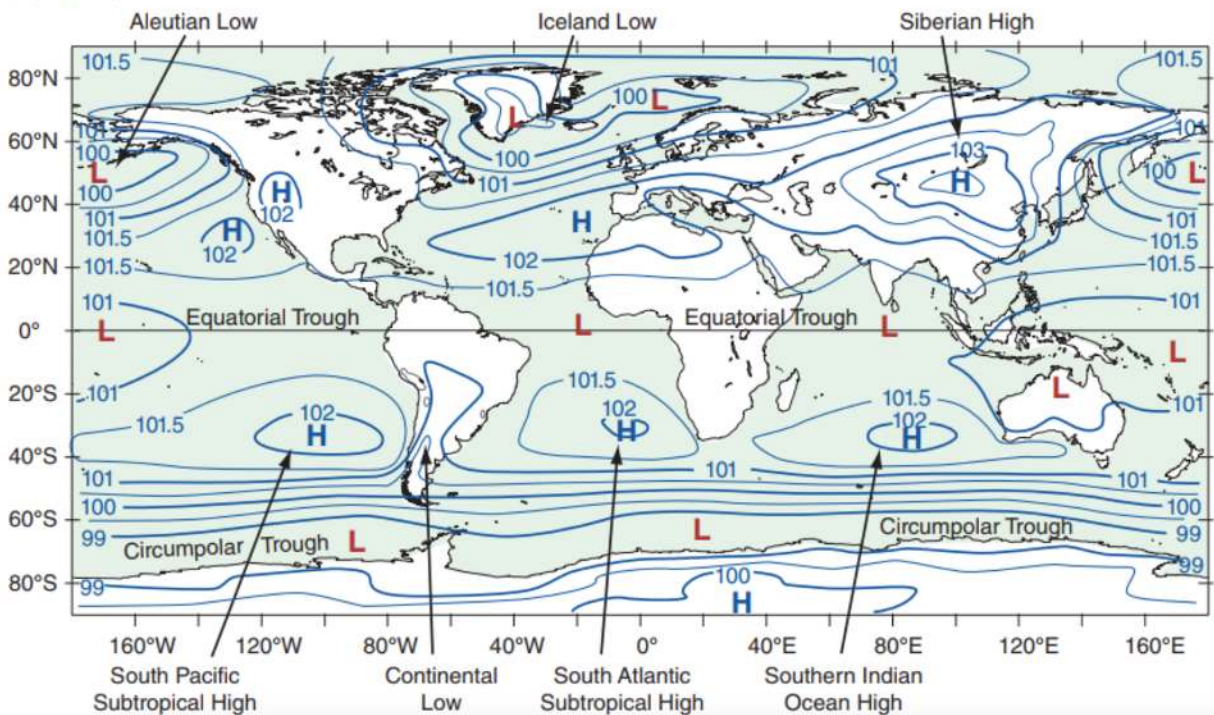


Average near surface winds ([CC BY-NC-SA 4.0](https://creativecommons.org/licenses/by-nc-sa/4.0/)).

How does the real world's average surface sea level pressure field compare with the above picture? When we add in the continents, ice masses, oceans, mountains, and forest, we get an average that looks something like the below two figures. The following maps show the mean sea-level pressure field for January and July, averaged from 1981 to 2010.

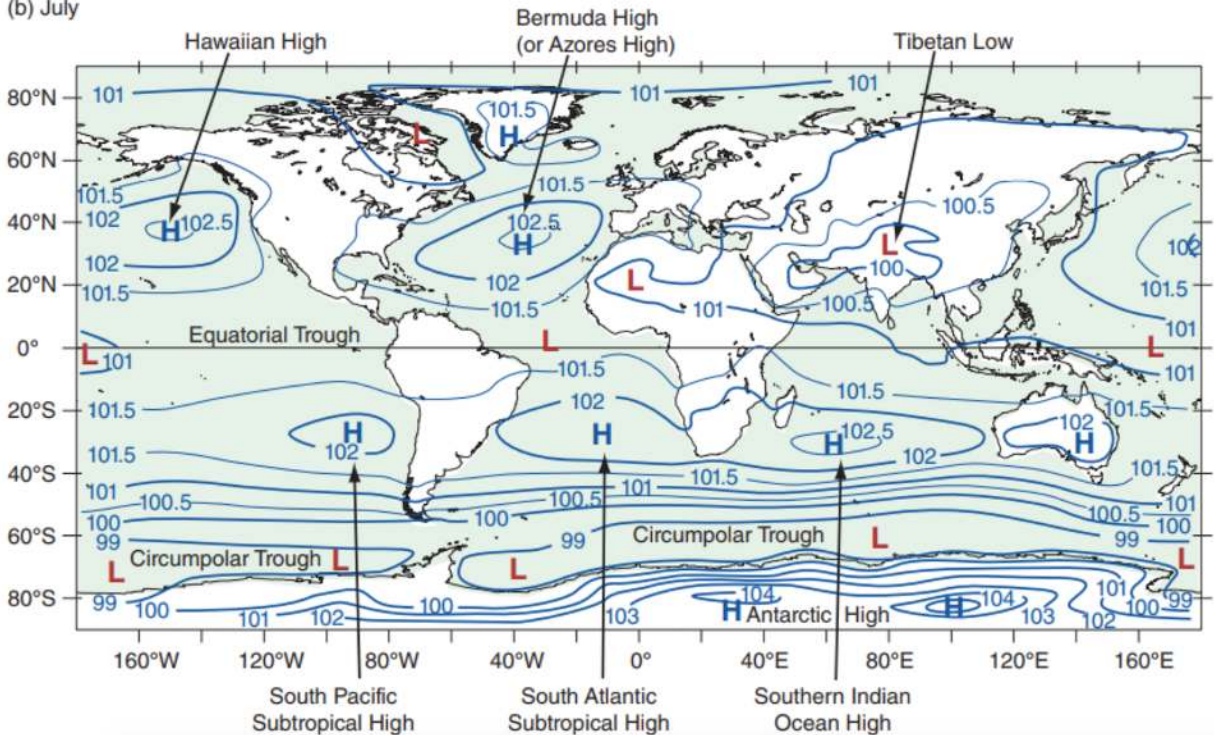
Looking at the two maps below, you may notice that there are some areas where low and high pressure systems seem to persist throughout the year – these are known as **semipermanent highs** and **semipermanent lows**. These include the **Bermuda-Azores High**, the **Pacific High**, the **Icelandic Low**, and the **Aleutian Low**.

(a) January



Map of semipermanent highs and lows during the month of January ([CC BY-NC-SA 4.0](#)).

(b) July

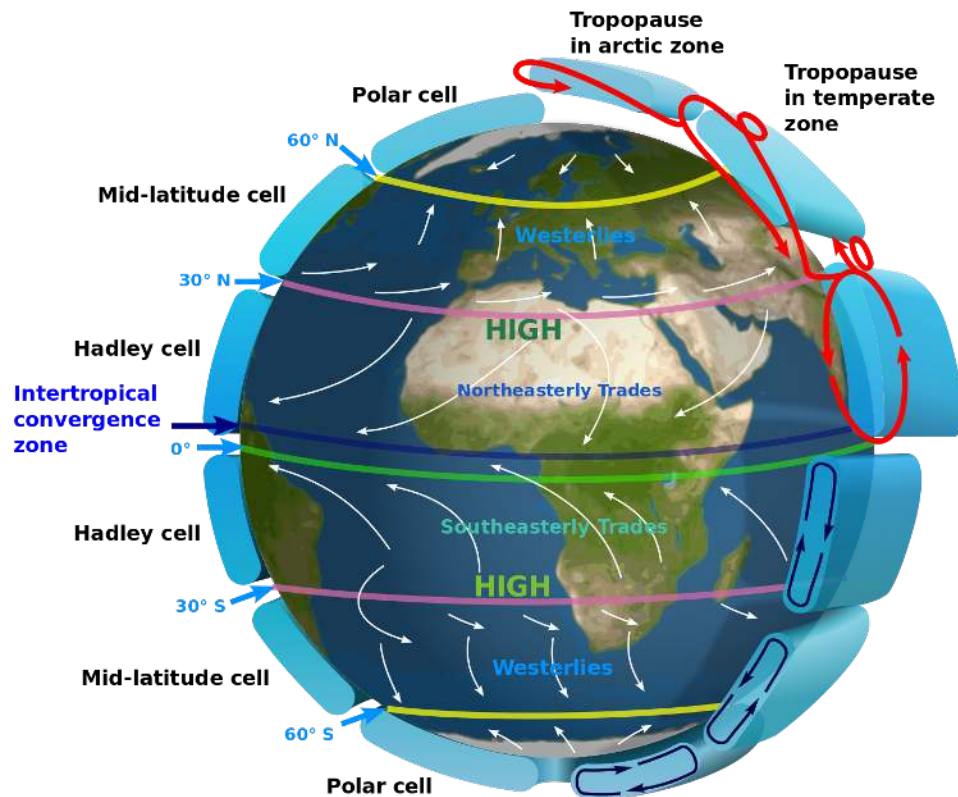


Map of semipermanent highs and lows during the month of July ([CC BY-NC-SA 4.0](#)).

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10.5: Global Surface Winds

The following graphic nicely illustrates all of the above phenomenon for global surface winds. The polar easterlies, mid-latitude westerlies, and tropical trade winds are all visible. Now with the continents and land masses added, we are able to see where on Earth these surface winds are observed.

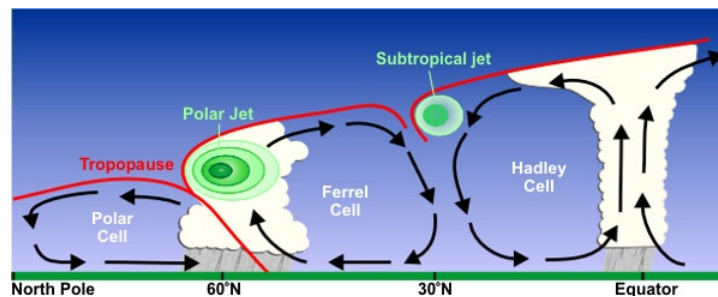


Circulation of Earth's atmosphere ([CC BY-SA 3.0](#)).

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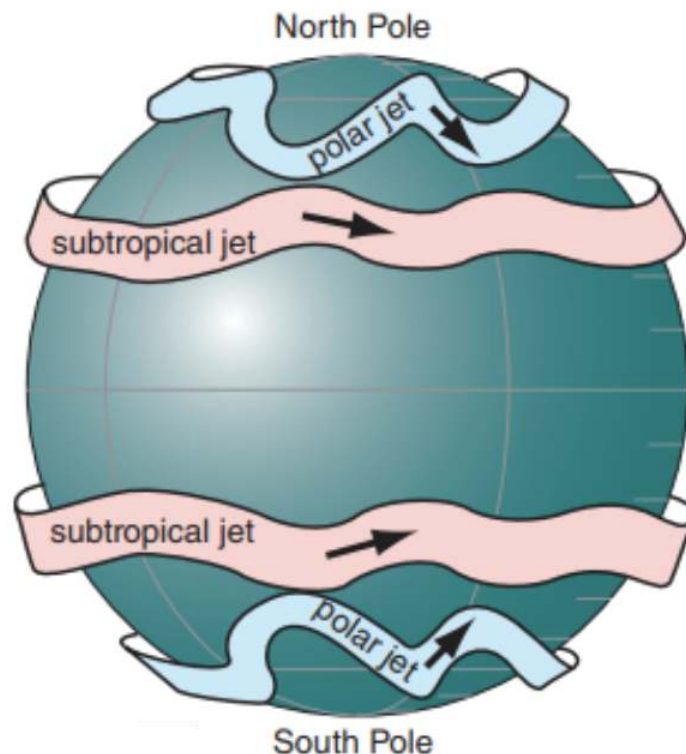
10.6: Jet streams

There is one final important piece of general circulation that deserves a discussion, and that is jet streams. In the below image you can see two jet streams: the subtropical jet and the polar jet. The figure shows the average position of the jet streams in the Northern Hemisphere in the winter, as well as their relation to the tropopause. The figure shows jet streams flowing from west to east. We can see that there are two jets located right under the tropopause. The **subtropical jet stream** is located near 30° latitude about 13 km up, above the tropical high. The **polar jet stream** is located near the polar front about 10 km up, near 50° to 60° latitude. The difference in height of these jets is due to their location at the tropopause, and the fact that the tropopause is found higher in tropical regions than in polar regions due to average layer temperature differences of the troposphere underneath. The troposphere is thinner in polar regions than in tropical regions due to colder, denser air at the poles.



Cross-section of the northern hemisphere circulation, and the positions of the polar and subtropical jet streams ([Public Domain](#)).

If the general circulation of the atmosphere is like a giant meandering river of air around the globe, then **jet streams** are swiftly flowing currents within that river. Jet streams are thousands of kilometers in length, and hundreds of kilometers in width. In the core of a jet stream (called a **jet streak**), wind speeds are often higher than 100 knots and are occasionally higher than 200 knots. The polar jet can sometimes merge with the subtropical jet if it sweeps southward enough, and it occasionally splits into two jet streams.



The position of the polar jet stream and the subtropical jet stream ([CC BY-NC-SA 4.0](#)).

The existence of jet streams is ultimately due to the energy imbalance between tropical and polar regions. How exactly do they form? As mentioned before, the polar front is a boundary between colder polar air and warmer subtropical air. Because of this, the strongest temperature gradient occurs along the polar frontal zone. This rapid change of temperature with distance also causes a rapid pressure change, due to the thermal wind effect (a vertical shear in the geostrophic wind caused by a horizontal temperature gradient). This strong pressure gradient across the polar front causes intense wind speeds that become the jet stream. The temperature contrast between north and south along the polar front is more intense during the winter than during the summer, so the polar jet is also stronger during the winter. During winter, the leading edge of the cold polar air pushes further south into subtropical areas. During the summer, the polar front retreats into higher latitudes and is weakened.

The subtropical jet stream tends to form just above the descending branch of the Hadley cell, at about 12 km altitude. Here, a boundary exists between warmer equatorial air and cooler air that has been cycled up and around the Ferrel cell from the polar front. This is sometimes referred to as the *subtropical front*, but it does not extend all the way to the surface. Here, the temperature gradient is strongest aloft near the tropopause, which induces a sharp pressure gradient and strong winds aloft as well.

One final thought for Chapter 11: isn't it fascinating that all of these global winds are caused by differential heating and the rotation of Earth? Everything is connected in one way or another and fits together like a puzzle.

Chapter 11: Questions to Consider

- 1.
2. What assumptions are made for a single-cell model of Earth's atmosphere?
3. Drag the terms to their correct position:
4. Where are jet streams located? Why do they differ in height?

Selected Practice Question Answers:

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CHAPTER OVERVIEW

11: Fronts and Airmasses

Alison Nugent and Shintaro Russell

Learning Objectives

By the end of this chapter, you should be able to:

- Describe what an “air mass” is and how it forms
- Name some of the main types of air masses
- Draw a vertical cross-section schematic of a warm and cold front
- Recognize the symbols for warm, cold, and occluded fronts
- Discuss the types of cloud patterns associated with warm and cold fronts
- Explain how differences in temperature (warm or cold fronts) or differences in humidity (dry lines) are important to weather

[11.1: Introduction](#)

[11.2: Air Masses](#)

[11.3: Surface Fronts](#)

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11.1: Introduction

We've already learned that areas of low and high pressure can be identified based on the isobars on a weather map. We've also learned that some regions on Earth typically have low pressure and while others typically have high pressure. In this chapter we'll learn that some regions of the atmosphere have similar air properties and are named by those properties as a collective mass of air or "air mass". High pressure systems, especially, are very common air masses.

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11.2: Air Masses

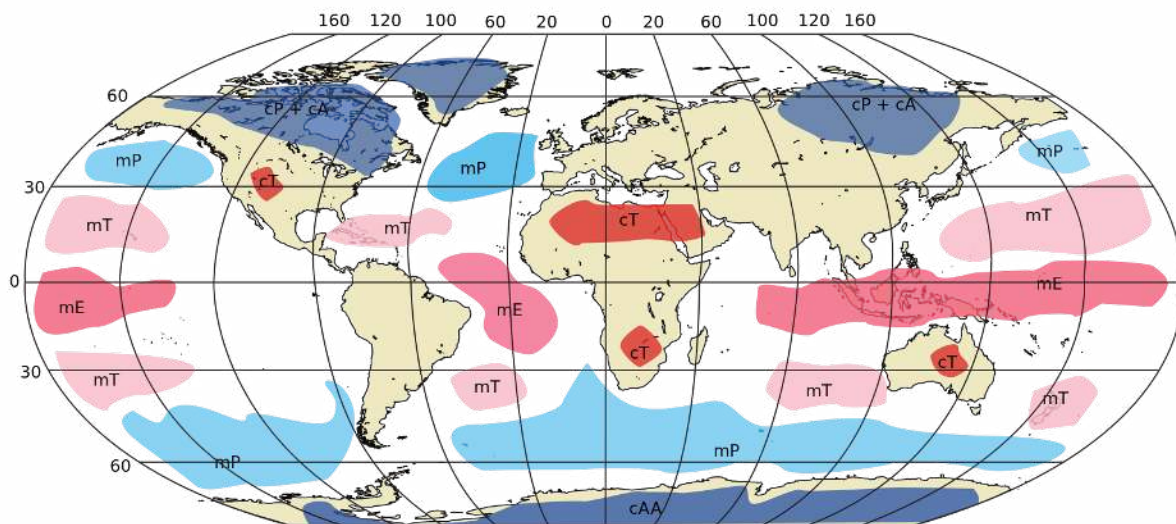
An air mass is an extensive body of air featuring generally similar temperature and moisture characteristics. They can extend thousands of square kilometers. Air masses are identified based on their temperature and humidity characteristics as well as their geographical region of origin. For example, if an air mass is dry and warm and originated from the tropical region over a continent, it would be called a continental Tropical (cT) air mass. If an air mass is humid and cold and originated over the ocean in the high latitudes, it would be called a maritime Polar (mP) air mass. The air mass name will always begin with a lower-case letter signifying either continental (c) or maritime (m) and a second capital letter signifying Equatorial (E), Tropical (T), Polar (P), Arctic (A), or Antarctic (AA).

Maritime air masses are humid air masses originating from oceans or large bodies of water. Continental air masses are dry air masses originating from land. Equatorial air masses are warm moist air masses originating from the equatorial region. Tropical air masses are warm air masses originating from the lower latitudes. Polar air masses are cold air masses originating from the upper latitudes. Arctic air masses are composed of extremely cold air that originated from the poles. Arctic air masses are even colder and drier than Polar air masses, and Antarctic air masses are even more extreme than Arctic air masses.

The common types of air masses are maritime Tropical (mT), maritime Polar (mP), continental Polar (cP), continental Tropical (cT), and continental Arctic (cA).

- Maritime Tropical (mT) air masses are warm, humid air masses originating from the oceans in the tropics.
- Maritime Polar (mP) air masses are cold, humid air masses originating from the oceans in the polar latitudes.
- Continental Polar (cP) air masses are cold, dry air masses originating from land regions in the polar latitudes.
- Continental Tropical (cT) air masses are hot, dry air masses originating from land in the tropics.
- Continental Arctic (cA) air masses are cold, dry air masses originating from the North Pole.
- Continental Antarctic (cAA) air masses are extremely cold and dry air masses originating from land at the South Pole.

See the below image to visualize where these different types of air masses typically originate.



Air masses and their location of origin ([Public Domain](#)).

Creation of Air Masses

Air masses develop when air is present over a surface for an *extended period of time*. This typically occurs in a high pressure system with light wind. The areas where air masses develop are called source regions. Air masses over warm surfaces usually develop faster than those over colder surfaces because there is weaker turbulence in the stable air over the cold surface. When air masses shift from their source regions, they change over time due to the surfaces and terrain over which the air masses flow.

Movement of Air Masses

Air masses do not remain over their source regions permanently. Slight changes in weather patterns may shift the air mass to a new location. As air masses move, two things can occur. First, as the air shifts over the different surface characteristics, the air mass begins changing. This process is called **air mass modification**. For example, an mP airmass that moves from the Pacific ocean over the mountains in the western continental US will typically dry as it crosses the mountains, rains out its moisture, and warms over the land surface until it becomes a cT airmass. The second thing that happens when air masses move is that they may collide with other air masses. When a collision occurs, the two air masses develop a boundary called a **front**.

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11.3: Surface Fronts

Surface fronts are the boundaries or transition zones between air masses at the Earth's surface. Changes in temperature, humidity, wind, pressure, visibility, as well as particular cloud and precipitation patterns are often observed at fronts. There are four main types of fronts.

1. Cold fronts
2. Warm fronts
3. Occluded fronts
4. Stationary fronts

Fronts are named based on the characteristics of the air mass that is replacing the prior air mass. For example, if a cold air mass is moving toward a warm air mass, the boundary between them will be called a cold front because the cold air is effectively replacing the warm air from the perspective of a stationary point on Earth's surface.

Fronts are usually associated with low pressure systems. Frontal boundaries on a map are labeled in the location where the temperature gradient of the front meets the Earth's surface.

Cold Fronts

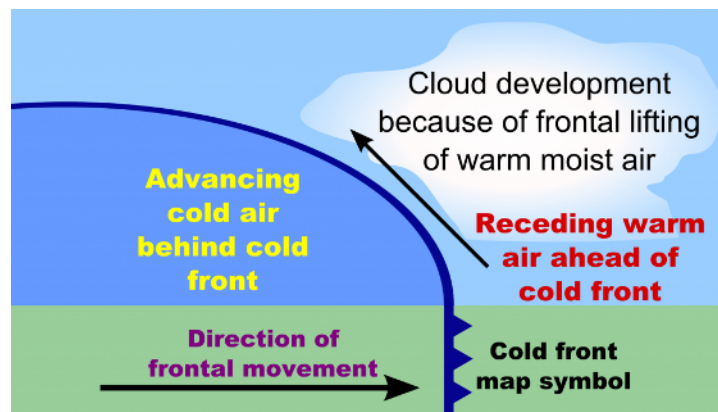


Diagram showing a vertical cross section through a cold front ([CC BY-SA 4.0](#)).

Cold fronts are a transition zone in which the advancing cold air mass replaces a retreating warmer air mass. On weather maps, cold fronts are drawn as blue lines with blue triangles pointing toward the warmer air mass, in the direction of frontal movement, as seen in the above image.

Because warm air is less dense than colder air, the cold air stays on the bottom and warm air is forced to rise above the advancing cold air. This forced lifting results in typical cloud patterns ahead of a cold front, which include cirrus and cirrostratus clouds. The number of cumulus clouds in the warm air mass increases as the frontal boundary approaches. Because the warm air mass is forced to rise, atmospheric instability occurs along the cold front and results in towering cumulus and cumulonimbus clouds, which may produce heavy rain and thunderstorms along the frontal boundary. During the passage of a cold front, the wind direction generally shifts from south or southwest (in the warm sector) to west or northwest (in the cold sector) in the northern hemisphere. After a cold front's passage, fair weather returns with the appearance of cumulus and stratocumulus clouds.

Warm Fronts

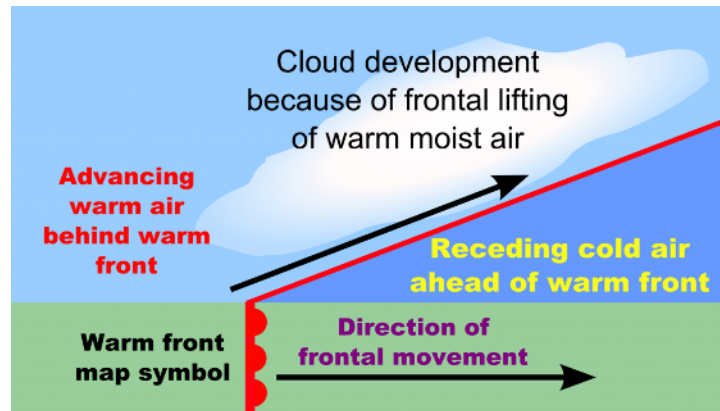


Diagram showing a vertical cross section through a warm front ([CC BY-SA 3.0](#)).

Warm fronts are a transition zone in which the advancing warm air mass replaces a retreating colder air mass. On weather maps, warm fronts are drawn as red lines with red semicircles pointing toward the colder air mass in the direction of the frontal movement.

Advancing warm air is forced to rise above the retreating cold dense air. Again, because of this forced lifting, typical cloud patterns are common ahead of a warm front. These include upper-level clouds such as cirrus and cirrostratus clouds before clouds thicken and lower-level clouds like altostratus, nimbostratus, and fog near the frontal boundary. Because the air mass is rising along the warm front, clouds form and steady precipitation may occur. During the passage of a warm front, wind direction shifts from east or southeast (in the cold sector) to southwest (in the warm sector) in the northern hemisphere. After a warm front's passage, cloud cover and precipitation decreases with only scattered cumulus clouds remaining.

Occluded Fronts

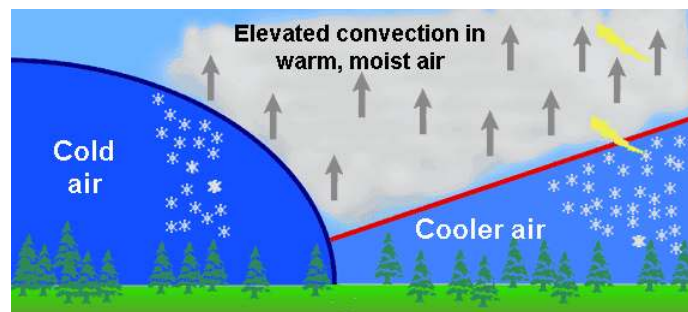


Diagram showing a vertical cross section through an occluded front ([Public Domain](#)).

Occluded fronts are a frontal boundary that forms when a cold front catches up to a warm front. Cold fronts move faster than warm fronts, so cold fronts can sometimes catch up to warm fronts, but not the other way around.

There are two types of occlusions: cold occlusions and warm occlusions. Cold occlusions (shown above) occur when the advancing air mass is colder than the retreating air mass. Warm occlusions occur when the advancing air mass is warmer than the retreating air mass. On weather maps, occluded fronts are drawn as purple lines with purple triangles and purple semi circles. You'll notice that this symbol is a combination of the cold front and warm front symbols, which isn't surprising because an occluded front is effectively a combination of the two.



Frontal symbol for an occluded frontal boundary ([Public Domain](#)).

Occluded fronts are the final stage of a frontal boundary because warm air above cool air is a stable scenario.

Stationary Fronts

Stationary fronts are a type of frontal system that are almost stationary with the winds flowing nearly parallel and from the opposite paths in each side separated by the front. On weather maps, stationary fronts are drawn as alternating blue and red lines with blue triangles pointing toward the warmer air mass and red semicircles pointing toward the colder air mass. This is the only scenario where the direction of the symbols does not indicate a direction of movement.



Frontal symbol for a stationary frontal boundary ([Public Domain](#)).

Other Frontal-like Features

Not all weather features have frontal characteristics, such as a trough and a dryline. While troughs feature a change in cloud and precipitation pattern, it lacks the sharp change in temperature and moisture as observed in fronts. Drylines are a boundary between warm, moist air and warm, dry air. Because both air masses are warm, a dryline cannot be classified as either a warm front or a cold front. Drylines commonly occur during spring and summer in southwestern United States, particularly in Texas. Warm, humid air from the Gulf of Mexico meets with the warm, dry air from the desert plateau. During the afternoon, convective clouds and even thunderstorms can develop in drylines as moist air rises over the denser dry air.

Chapter 12: Questions to Consider

1. What is an air mass? What are some different types of air masses and where do they originate?
2. Label the cold front cross-section:
3. Label the warm front cross-section:
4. Match the front name with the correct symbology:
5. What cloud and precipitation patterns are typically associated with warm and cold frontal passage?

Selected Practice Question Answers:

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CHAPTER OVERVIEW

12: Extratropical Cyclones

Alison Nugent and David DeCou

Learning Objectives

By the end of this chapter, you should be able to:

- Describe how cyclogenesis occurs
- Identify areas on a map where mid-latitude cyclones are common, and explain why they move where they do
- Sketch the frontal systems involved in a mid-latitude cyclone
- Understand the hazards associated with mid-latitude cyclones
- Discuss the relationship between sea level pressure, high and low pressure systems, air columns and mass budgets as a closed system

[12.1: Introduction](#)

[12.2: Mid-latitude Frontal Cyclones](#)

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12.1: Introduction

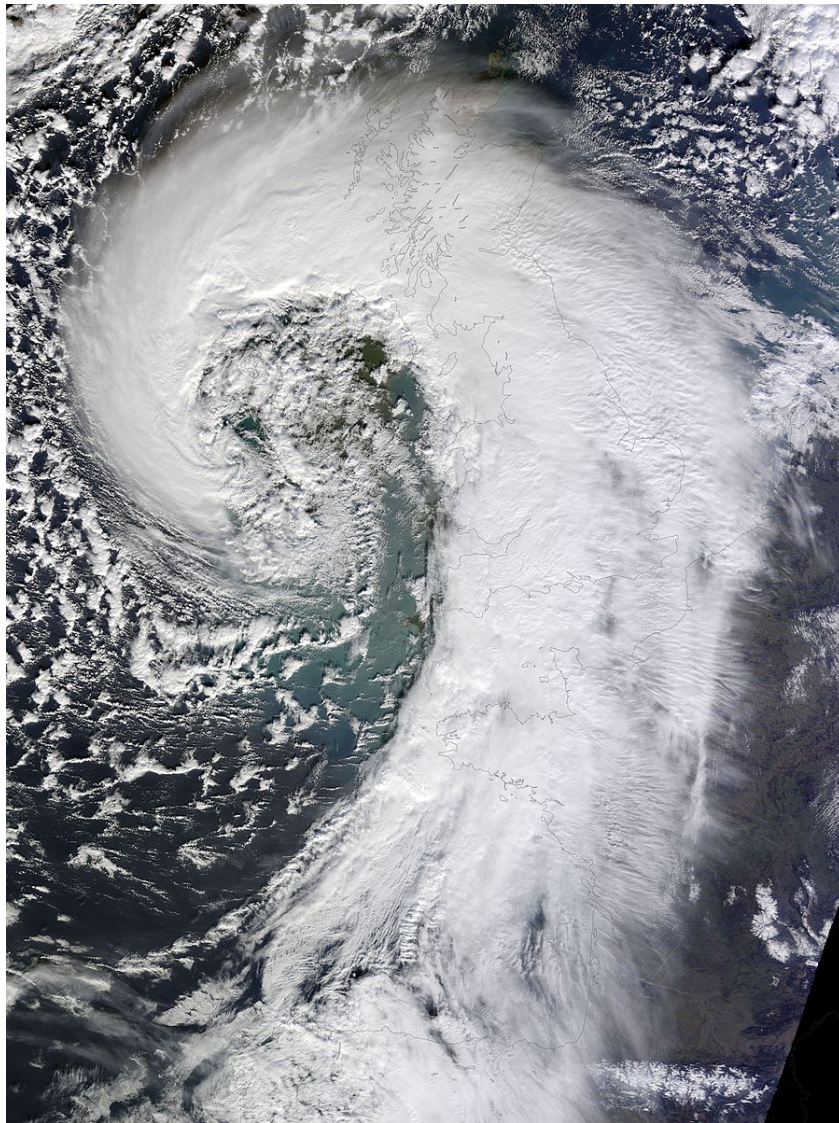


Satellite image of a mid-latitude cyclone over North America ([Public Domain](#)).

For well over a century, forecasters have been aware that areas of falling barometric pressures are often accompanied by precipitation and strong winds. However, it wasn't until the early 1900's that atmospheric scientists began piecing together a more complete picture of how low pressure systems develop, as well as the weather associated with them.

Recall that a cyclone is an area of low pressure, around which winds blow counterclockwise in the Northern Hemisphere and clockwise in the Southern Hemisphere. This is due to the fact that winds blow from high to low pressure, but are deflected by the Coriolis force (perpendicular to the right of the motion vector in the Northern Hemisphere, left in the Southern Hemisphere). The focus of this chapter is cyclonic storm systems that form in the mid-to-high latitudes outside of the tropics. These storm systems are either called mid-latitude frontal cyclones, extratropical cyclones, wave cyclones, or simply frontal cyclones. Tropical cyclones will be the focus of a later chapter.

Shortly after World War I, Vilhelm Bjerknes, Jakob Bjerknes, Halvor Solberg, and Tor Bergeron published their *Norwegian Cyclone model*. This model proposed a life cycle for the development of mid-latitude cyclones, and was mostly based on surface observations. It became known as the **Polar Front Theory** of a developing wave cyclone. It was eventually modified and today provides a way to describe the structure, weather, and evolution of a moving cyclonic storm system in the mid-latitudes. First we will look at how a mid-latitude cyclone develops at the surface, and then we will look at how the surface evolution is affected by the winds aloft.



Satellite image of an extratropical cyclone over the UK ([CC BY 2.0](#)).

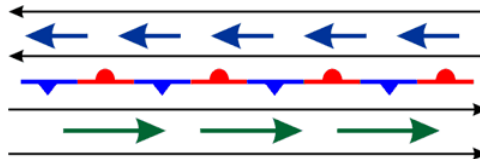
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12.2: Mid-latitude Frontal Cyclones

Following the Norwegian model, the development of a mid-latitude cyclone begins along the polar front. Recall from Chapter 11 that the polar front separates cold polar air from warmer subtropical air at around 60° latitude. It is a semi-continuous boundary and mid-latitude cyclones form and move along it as a series of waves. For this reason, a developing storm is sometimes referred to as a wave cyclone.

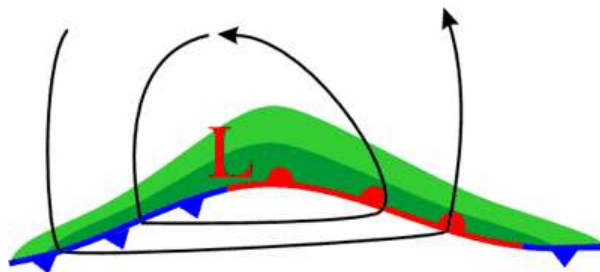
Cyclogenesis and Life Cycle

This figure below shows a portion of the polar front as a stationary front, with cold air to the north and warmer air to the south flowing parallel to the front in opposite directions. These winds moving in opposite directions set up rotation, similar to how a pen will turn if you place it between your hands and move them in opposite directions.



Part 1 of cyclogenesis: a stationary front with opposite moving winds on either side ([Public Domain](#)).

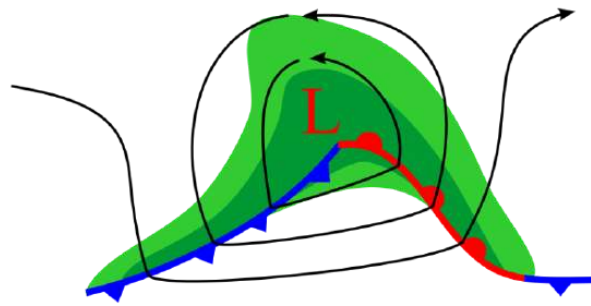
Under the right conditions, a **frontal wave** will begin to form along the front, with a cold front pushing southward and a warm front moving northward. The lowest pressure lies at the junction of the two fronts. The southward-moving cold front pushes warmer, less dense air upward, while the warm front overruns and moves over the colder air ahead of it. This creates rising motion in the column, and a narrow band of precipitation forms. The surface low pressure system is steered by winds aloft, typically moving eastward or northeastward, and it gradually becomes a fully-developed mature cyclone 12 to 24 hours after its incipient stage.



Part 2 of cyclogenesis: the formation of a frontal wave ([Public Domain](#)).

The central pressure lowers and the pressure gradient increases, causing a stronger cyclonic (counterclockwise) flow inward toward the low's center. The precipitation band widens ahead of the warm front, and narrows ahead of the cold front. The region of warmer air between the cold and warm fronts here is called the **warm sector**. Here, the weather is generally partly cloudy, with scattered showers possible if the air is conditionally unstable.

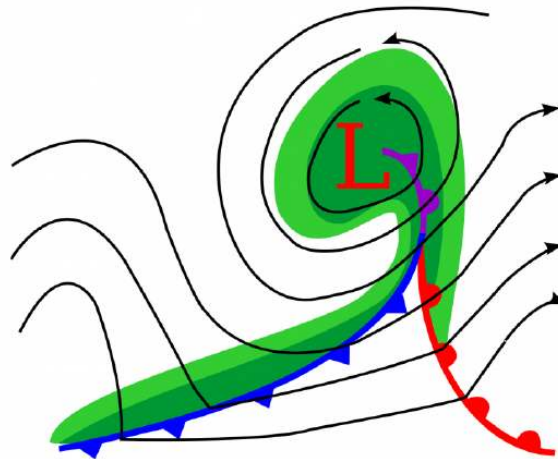
Where does the mid-latitude cyclone get its energy from? As the warmer and colder air masses attempt to regain equilibrium, warm air rises over the colder air, which transforms potential energy into kinetic (motion) energy. As warmer air rises, it condenses into clouds, which release latent heat energy into the system. Near the surface, winds converge inward toward the low's center. Wind speeds may increase as a result of a stronger pressure gradient force near the center, increasing the kinetic energy in the system.



Part 3 of cyclogenesis: a newly developed cyclone ([Public Domain](#)).

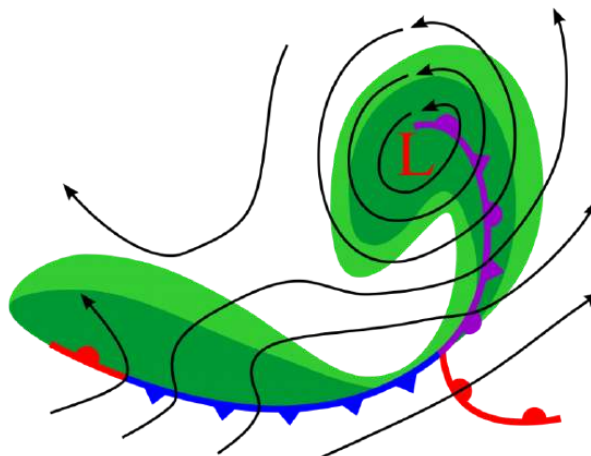
As the cyclone moves eastward, the central pressure continues to decrease and winds increase during its mature stage. The faster-moving cold front closes in on the warm front, decreasing the size of the warm sector. In the Norwegian cyclone model, the cold front overtakes the warm front and the cyclone becomes occluded. Cloud cover and precipitation cover a wide area and the storm is usually most intense at this stage.

The point where the cold front, warm front, and occluded front intersect is called the *triple-point*. Occasionally, a secondary low may form at this triple point, move eastward, and intensify into another cyclone.



Part 4 of cyclogenesis: formation of an occluded front and triple point ([Public Domain](#)).

Eventually, as occlusion advances, the low pressure center will begin to dissipate, because cold air exists on both sides of the occluded front. The sector of warm, rising air is removed from the center of the storm, so the storm gets cut off from its primary energy supply. Eventually the old storm dies out and gradually disappears.



Part 5 of cyclogenesis: continued formation of an occluded front and dying out of the cyclone ([Public Domain](#)).

This sequence of a developing mid-latitude cyclone is similar to a whirling, spinning eddy in a river that forms behind a stick or log, moves along with the river, and quickly disappears further downstream. This entire life cycle can last from several days to a little more than a week.

Cyclones in various stages of development can be seen all at once along the polar front—this succession of storms is known as a cyclone “family”. As mentioned before, some cyclones form from dying previous cyclones and become a part of the succession.

This polar front model of development for a mid-latitude cyclone is rather simplified and, in fact, very few storms follow this model exactly. However, it is a good foundation for understanding storm structure.

Storm Tracks

The development of a mid-latitude cyclone is a process called **cyclogenesis**. Certain regions in North America are more favorable for cyclogenesis, including the eastern slopes of mountain ranges like the Rockies and Sierra Nevada, the Atlantic Ocean off the Carolina Coast, and the Gulf of Mexico. When air flows westward across a north-south extending mountain range, the air on the leeward (downwind) side tends to have cyclonic curvature, which adds to the development of a cyclone. This is called **lee cyclogenesis**, and cyclones that are a result of this are often called **lee-side lows/cyclones**.

Cyclones may also develop near Cape Hatteras, North Carolina, where warm moist air from the Gulf Stream can increase the north-south air mass temperature/moisture contrast to the point where cyclogenesis might occur. These cyclones are called **northeasters** (or nor’easters) and normally move northeast along the Atlantic Coast. These storms can bring heavy rain or snow and high winds to areas along the East Coast.

Typical cyclone storm tracks are named after the region in which they form, like the *Hatteras low*, *Alberta Clipper*, or *Colorado low*. Alberta clippers and Colorado lows form or re-develop on the lee-side of the Rockies. Mid-latitude cyclones always move toward the east due to the prevailing westerlies.

What influences the strength of a mid-latitude cyclone, and determines how long it will persist? There are some surface conditions that influence cyclogenesis, but the real key to mid-latitude cyclone development lies in the winds aloft. How are mid-latitude cyclones influenced by upper-level flow?

Mid-latitude Cyclone in Three Dimensions

Developing surface lows are usually more intense with height and appear on upper-level charts as a trough or a closed low. However, the low in the upper-levels usually exists to the west of the surface low (again, in the Northern Hemisphere). This is a necessary condition for a low pressure system to continue to develop and intensify. If the upper-level low were directly over the surface low, the surface low would quickly dissipate. This is because winds converge inward toward the low, but only at the surface. This convergence at the surface causes the air mass to “pile up” and air density to increase just above the surface low. The increase in air mass causes surface pressures to rise, and the low fills in and dissipates. How then do cyclones intensify and develop? The air that piles up at the surface must have an “exit path” out of the column so that the surface air pressure can continue to decrease and the cyclone can strengthen. The vertical structure of the atmosphere must allow for air to rise out of a surface low pressure.

The following figure shows an idealized model of the vertical structure of a cyclone and anticyclone in the Northern Hemisphere. A surface low and a surface high are accompanied by an upper level trough and ridge respectively.

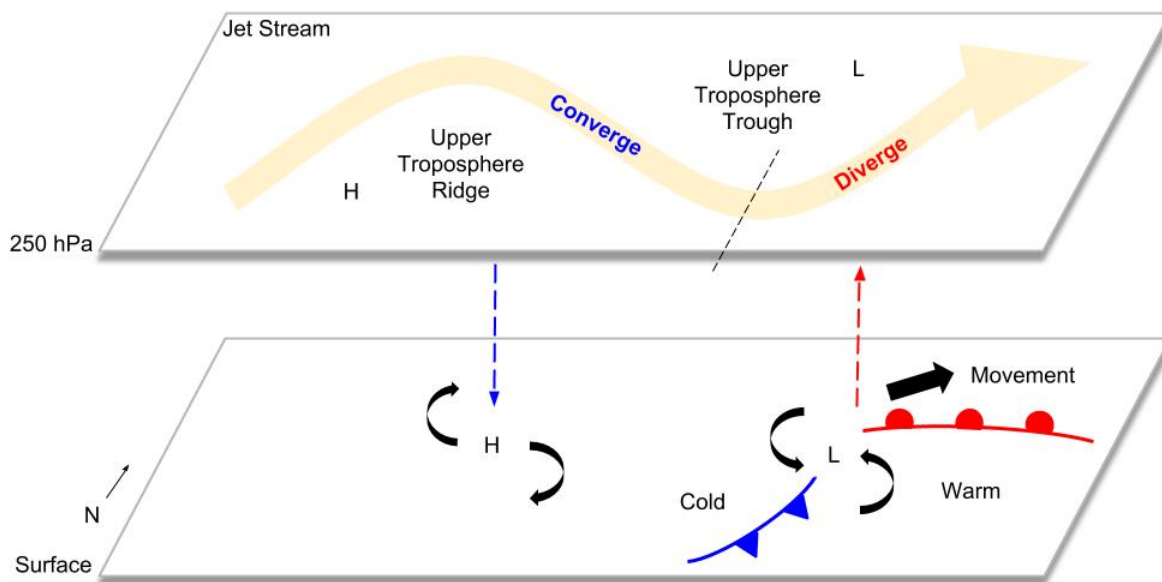


Diagram of the structure of a cyclone and anticyclone accompanied by a trough and ridge (Image Created by Britt Seifert).

On the right hand side is a Northern Hemisphere frontal cyclone with a warm and cold front. The cold air behind the cold front at the surface also extends upward aloft. Recall that the layer between two pressure surfaces is thinner when the air temperature of the layer is cold (more dense), and thicker when the layer is warm (less dense). When pressure levels are packed closer together, pressure decreases more rapidly with height in a column of cold air. Because of this, low pressure is found aloft a body of cold air, just as you find behind a cold front because the constant pressure surfaces are squeezed closer to the Earth's surface. Thus an upper low is often found in the cold air aloft to the west of, or behind, the surface low. The surface low tilts toward the northwest moving up from the surface. Directly above the surface low, the airflow spreads out and diverges. This allows the converging surface air to rise and flow out of the air column at the tropopause, reinforcing vertical motion. When the divergence in the upper levels is stronger than convergence at the surface, surface pressures will lower further, and the low will intensify and deepen. Put another way: when air exits the column more rapidly aloft than it enters the column at the surface, then the amount of air in the column will reduce, the surface pressure will lower, and the cyclone will intensify.

Notice that there is convergence directly aloft of the high pressure system. Here, the air mass increases aloft and piles up, while air flows clockwise (in the Northern Hemisphere) and out of the anticyclone at the surface. If air were able to flow freely out of the anticyclone, the air pressure would rapidly drop and the anticyclone would dissipate. Therefore, to maintain or strengthen the high pressure system, air has to continually be added to the anticyclone. This happens when there is convergence above a surface high. The air that piles up aloft sinks in the column increasing surface pressure. If the convergence aloft is stronger than the divergence at the surface (more air is added than is removed), then the surface pressure will increase.

Winds at the 500-mb pressure level tend to steer surface low and high pressure systems. Generally speaking, surface storm systems tend to travel at about 16 knots in summer, and roughly 27 knots in winter. This is due to stronger jets and upper-level flow in the winter, a result of stronger north-south temperature differences.

Redistribution of Heat

Mid-latitude frontal cyclones are both a vital part of global circulation and a result of global circulation. They're also an important pattern in the climatology of regions in the mid-latitudes.

The temperature gradients that cause frontal cyclones form as a result of the colliding surface air from the polar and Ferrel cells. The strong temperature gradient with cold air from the polar region and warm air from the tropics is the energy source that drives the frontal storms. On the flip side of the same token, frontal cyclones are a huge contributor to the redistribution of heat globally. Warm air moving poleward in a warm front and cold air moving equatorward in a cold front are later stabilized after the temperature gradient balances itself by forcing the cold air aloft (occluded front). Frontal cyclones are thus both a result of and a contributor to global circulation and the redistribution of heat from the equator to the poles.

The beginning of the chapter showed images of these storm systems extending over scales the size of continents and land-masses. It also described how these storm systems last from days to over a week. Instabilities along the polar front are always growing and dying, and passing over fixed points on Earth. The point is that if you live somewhere along the storm track in the Northern or Southern hemisphere, in the wintertime, these storm systems dictate your weather. For example, if you live in Boston, Massachusetts, your winter weather may look something like this: a few days of warm clear weather, a quick change (passing cold front) resulting in a large drop in temperatures and some heavy rain, then cold dry weather for a few days. This cold weather then transitions slowly to warm by some light rain and warming temperatures (warm front). While variable, this pattern repeats itself week after week. Depending on the stage of the frontal storm as it passes over you, it may be more or less severe, and you may receive more or less rain, snow, or other wintery weather. The precipitation and temperature variations resulting from frontal cyclones are an important part of the climatology of mid-latitude weather.

Chapter 13: Questions to Consider

1. Which front do mid-latitude cyclones form and move along?
2. How long does it take for a cyclone to fully develop?
3. What is the point where the cold front, warm front, and occluded front intersect called?
4. True or False:
5. Put the steps of cyclogenesis in the correct order from 1 to 5:

Selected Practice Question Answers:

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CHAPTER OVERVIEW

13: Thunderstorm Fundamentals

Alison Nugent and Shintaro Russell

Learning Objectives

By the end of this chapter, you should be able to:

- Draw a diagram and label parts of a thunderstorm
- Recognize that thunderstorms are sometimes part of a mid-latitude cyclone
- Describe the importance of updraft separation from the downdraft and precipitation portion of a thunderstorm
- Identify a few types of thunderstorms (airmass, MCC, squall line, supercell, etc.)
- Discuss a few favorable conditions for thunderstorm formation
- Describe the importance of wind shear for thunderstorms
- Identify LFC and EL on a skew-T as well as CAPE and CIN and connect them to thunderstorm characteristics
- Compute updraft velocity from CAPE

[13.1: Introduction](#)

[13.2: Thunderstorms](#)

[13.3: Atmospheric Instability and Thunderstorms New Page](#)

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13.1: Introduction



Lightning strikes the water from a thunderstorm in Northern Sicily ([CC BY-SA 2.0](#)).

Thunderstorms are deep convective clouds that have a large vertical extent all the way from the boundary layer to the tropopause. Thunderstorms often bring a variety of severe weather such as heavy rain, hail, lightning, damaging winds, and, occasionally, tornados. Thunderstorms are sometimes thought to be synonymous with cumulonimbus clouds, though not all cumulonimbus clouds are thunderstorms.

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13.2: Thunderstorms

The primary energy that drives thunderstorms is the conversion of moist air into clouds and precipitation, which releases significant amounts of latent heat in the condensation process. Over land, thunderstorms occur most frequently in the late afternoon and early evening. Storms are most likely to occur soon after the warmest surface temperatures, which helps to give warm air parcels the initial buoyancy they need to begin to rise. Thunderstorms can last anywhere from less than an hour to more than 12 hours. Because of their rapid growth, relatively small (mesoscale) size, and sensitivity to environmental conditions, it is difficult to predict the exact time and location of a storm. Typically a forecast will say there is a chance of thunderstorms over an area—meaning that although conditions are conducive for a storm, the forecasters don't know exactly where or when it will occur.

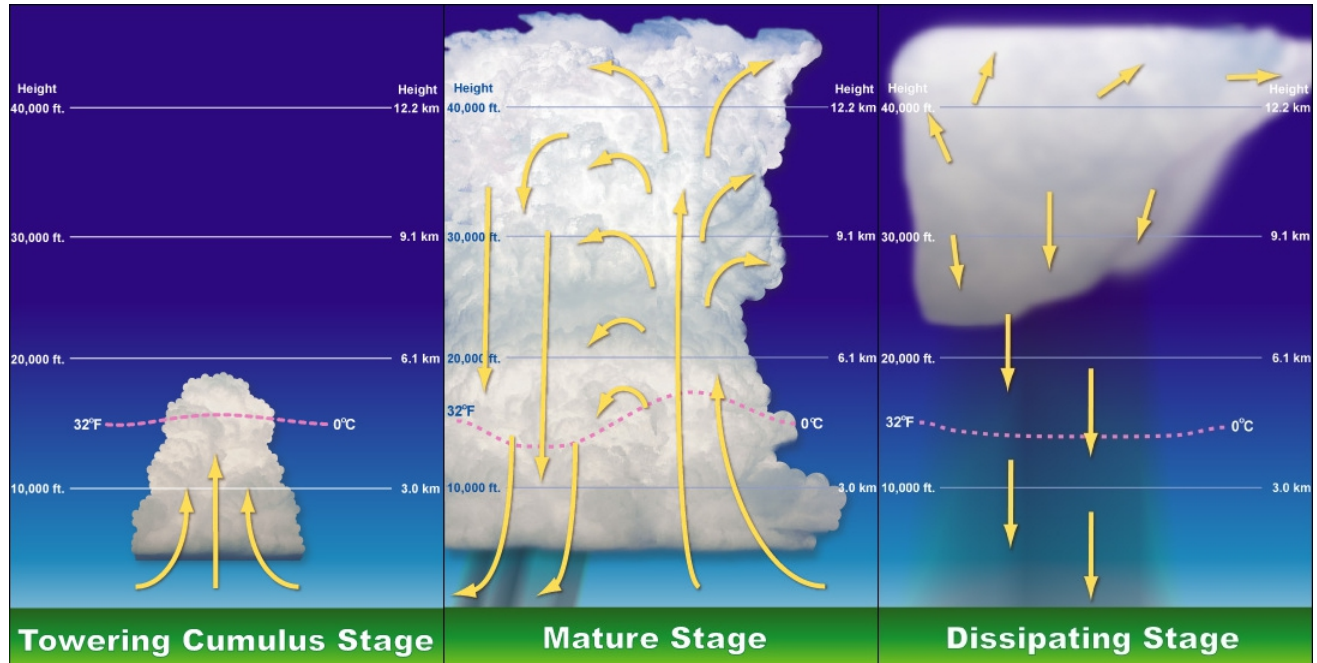
Thunderstorm Life Cycle

The three main stages of a thunderstorm are as follows:

1. Developing stage
2. Mature stage
3. Dissipating stage

In the **developing stage**, rising air called “**updrafts**” are dominant in this stage and towering cumulus clouds form. This type of cloud is also known as a cumulus congestus. Next is the **mature stage** where the towering cumulus becomes a cumulonimbus cloud featuring both updrafts and “**downdrafts**” or sinking air. Downdrafts develop as a result of water droplets falling to the ground as precipitation (e.g., snow, rain, hail). The falling drops drag air downward as they fall, creating a downward current of air in the cumulonimbus cloud. Finally, in the **dissipating stage** downdrafts are dominant and cut off the updrafts needed for a cumulonimbus cloud to form and sustain itself. As a result, the cloud dissipates.

If a cumulonimbus cloud has thunder and lightning, it is called a **thunderstorm**. Lightning is caused by a charge separation in clouds. For this charge separation to occur, a cloud must have ice crystals inside of it, which is why thunderstorms are always mixed-phase or ice-phase clouds.



Formation and dissipation of a thunderstorm ([Public Domain](#)).

Airmass Thunderstorm

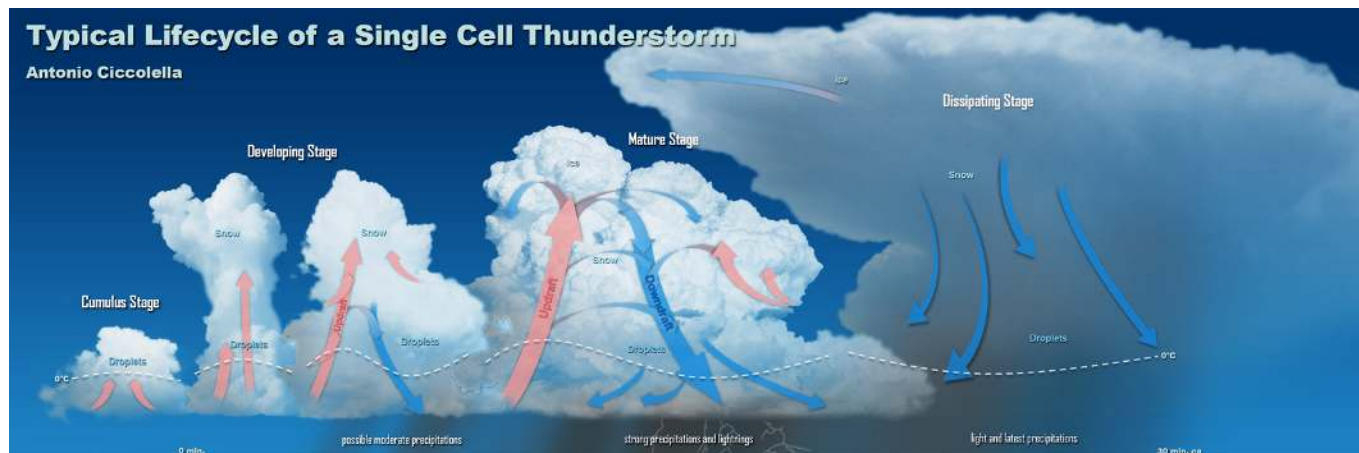
The thunderstorm described above is called an airmass thunderstorm. In general, the ingredients necessary for a thunderstorm to occur are **warm moist air**, some type of **instability**, and a **trigger** that can initiate the storm system. Airmass thunderstorms

typically occur in environments with similar properties and little wind shear. Airmass thunderstorms are the most benign types of thunderstorms and typically short-lived, lasting less than a few hours.

The reason airmass thunderstorms are so short-lived and benign is because the storm shuts off its ability to maintain itself. This is due to two primary reasons:

1. Below cloud base the falling rain evaporates, cooling the air in the boundary layer. The storm deprives itself of the heat and moisture in the boundary-layer air. Without the warm buoyant air to drive further convection, the storm loses its strength and dissipates.
2. Rain falls within the updraft of the storm, reducing its strength, again causing dissipation.

Another diagram of an airmass thunderstorm or single cell thunderstorm is shown below. The initial small cumulus cloud grows to a cumulus congestus, which has mostly updrafts. It then grows further into a mature cumulonimbus cloud with well defined updrafts and downdrafts. The final stage is the dissipating stage, which has mostly downdrafts.



Another diagram showing the typical cycle of a thunderstorm ([CC BY-SA 4.0](#)).

The trigger for a thunderstorm can be a warm humid air mass heated from the bottom by daytime solar heating or forced lifting by terrain (orographic thunderstorms). Collision of air masses can be another trigger, as you might find along a warm front, cold front, or dryline.

Thunderstorm Features

The main features of a mature thunderstorm includes updrafts, downdrafts, overshooting tops, and an anvil. An **overshooting top** occurs at the top of the thunderstorm. Overshooting tops are dome-shaped as a result of strong updrafts in intense thunderstorms reaching the tropopause and pushing upward into the stable region. However, ultimately, because the tropopause is so stable, the air gets pushed back downward and spread laterally. This lateral spreading forms the cloud **anvil** at the top of the thunderstorm along the tropopause boundary.

Severe Thunderstorm

While basic airmass thunderstorms typically consist of one “cell”, severe storms often contain multiple cells, consisting of more than one rising thermal connected with the same storm system. In a multicell storm, each cell is typically at a different life cycle stage and they continue on one from another.

In order for a thunderstorm to become severe, one important additional ingredient is necessary. In addition to warm moist air, some sort of instability, and a trigger, severe thunderstorms need **wind shear**. Wind shear is defined as a change in wind speed or direction with altitude. The reason wind shear is important for severe storms is easy; wind shear helps to tilt a storm such that the updraft and downdraft are displaced from one another. Without the competition between upward moving air and downward moving air, severe thunderstorms can strengthen and last much much longer than airmass thunderstorms. Severe thunderstorms come in many different forms. A few will be discussed below.

Severe Thunderstorm Types

The three main types of thunderstorms include squall lines, mesoscale convective complexes (MCCs), and supercells.

Squall Line

A **squall line** is a linear propagating severe thunderstorm. Squall lines consist of a long, narrow line of merged thunderstorms, often triggered by a cold front, dry line, or gust front. Squall lines can be many hundreds of kilometers long, but their width is usually much shorter, between 15 and 100 km. They can sustain themselves from several hours to several days.

A squall line continues propagating with the help of a gust front. When rain falls into the drier environment below the cloud base, it evaporates, cooling the air. This cool air continues to move downward, sometimes with strong force and reinforcement from a continuous downdraft. When the cool downdraft hits the surface of Earth, it spreads laterally, literally the same as a density current. Because it is relatively cool, it acts like a mini cold front, pushing warmer air up over it as it moves along. This upward push can initiate convection, and continue fueling the larger storm system.

Mesoscale Convective Complex

A mesoscale convective complex (MCC) is a type of severe storm that has a cloud shield (anvil) with a diameter of at least 350 km, elliptical or circular shape, and lasts between 6 and 12 hours. MCCs are huge storms that occur multiple times per year especially in the central United States. They are truly impressive masses of convection.

Supercell

A **supercell** is a special type of severe storm that has rotation. Supercells form when the environment has **directional wind shear** causing updrafts within a cloud to gain vorticity as they rise. While all severe storms have a structure that separates the updraft from the downdraft, supercell storms have a very clearly defined structure, which allows them to sometimes form tornados. As shown in the below image, supercells have additional features like a **wall cloud** and a **flanking line**. The wall cloud exists beneath the cloud base and near the border between the downdraft of cold air and the lower-level inflow of warm humid air. The flanking line is the region of towering cumulus that indicates extensive updrafts ahead of intense thunderstorms.

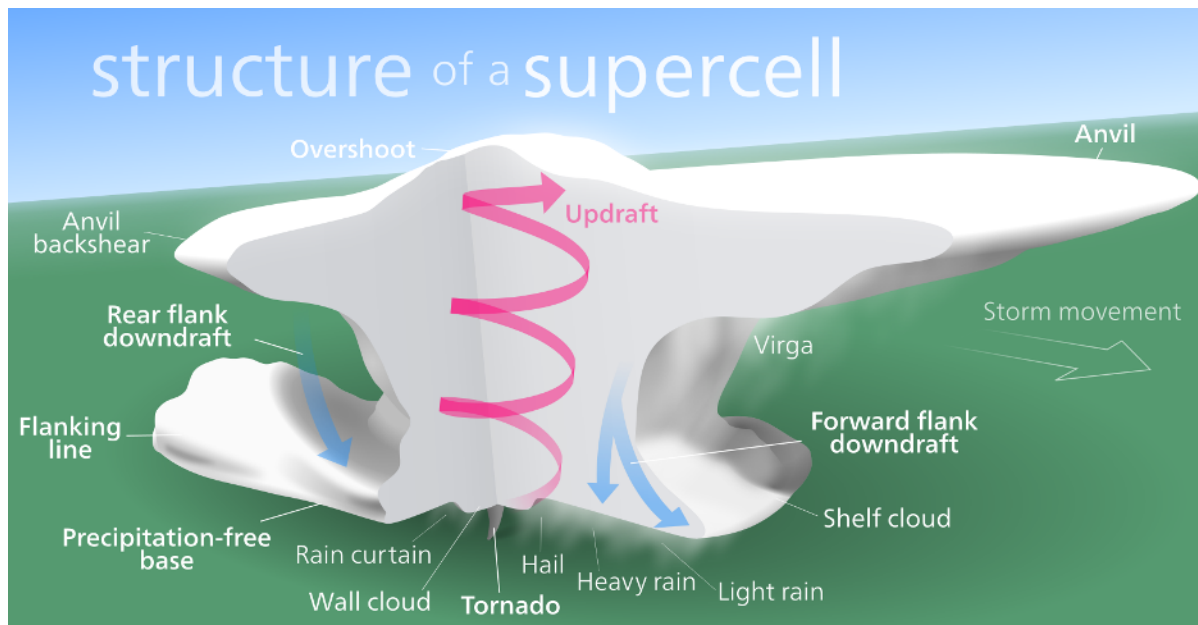
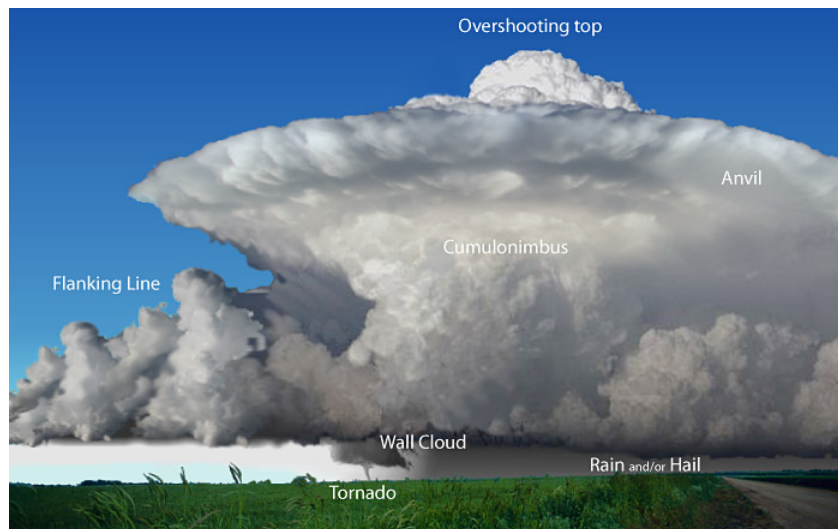


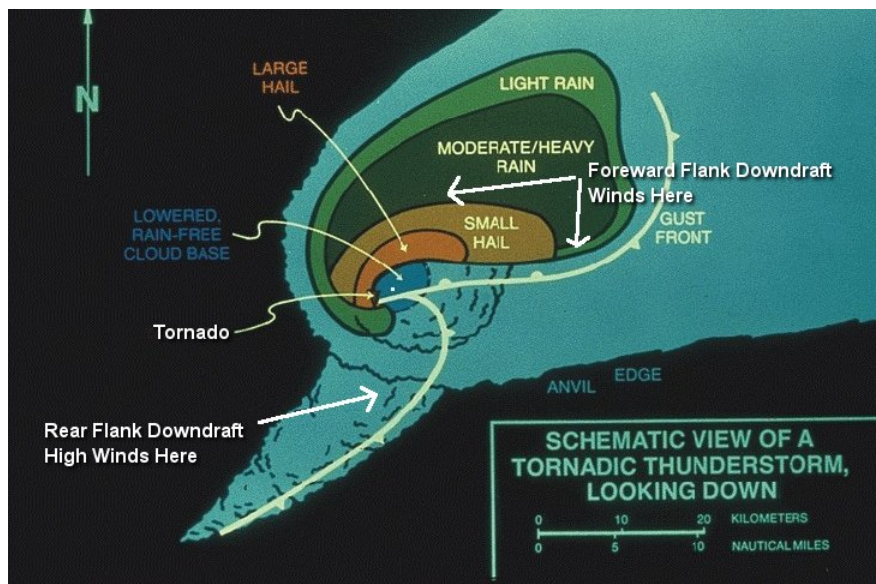
Diagram of a supercell thunderstorm with an anvil shape (CC BY-SA 3.0).

Here is a diagrammed image of an actual supercell thunderstorm that created a tornado. Remember that supercells are rotating storm systems that can produce tornados, but they don't always. In fact, very few supercells produce tornados, but the ones that do get the most attention!



An image of a supercell cumulonimbus cloud ([Public Domain](#)).

Here is one last image showing the structure of a supercell from the top view. The image shows the location of the gust front, the location of various precipitation types, and the location of downdrafts within the storm system.



A top view diagram of a supercell thunderstorm ([Public Domain](#)).

Thunderstorm Formation

Let's review. The ingredients needed for thunderstorm formation include high humidity, conditional instability, and a trigger that initiates rising air. A symmetric short-lived storm is called an airmass thunderstorm. When we add wind shear to an airmass thunderstorm, a severe thunderstorm can result. We discussed three types of severe thunderstorms including squall lines, MCCs, and supercells. Supercells are a type of severe storm with rotation resulting from directional wind shear in the environment.

High humidity in the atmospheric boundary layer is required for thunderstorms to occur. When water vapor condenses, latent heat is released. Latent heat is the primary energy source for thunderstorms. The higher the humidity, the more latent heat is released and the stronger the thunderstorm becomes.

Instability is necessary for convection to occur. The best case scenario is conditional instability, which often occurs when cold air lies above warm moist air capped by a temperature inversion. Cold air in the upper troposphere gives greater buoyancy to the updraft of warmer air from below and, therefore, a greater chance for strong thunderstorms to occur.

Wind shear is the change in wind speed and or wind direction with height. In an environment with wind but without wind shear, thunderstorms would last between 15 minutes and 1 hour as the thunderstorm and boundary-layer air would remain together. In this scenario, an airmass thunderstorm dissipates after it deprives itself of heat and moisture in the boundary-layer air. Strong wind shear pushes thunderstorms away from the depleted boundary-layer air and into areas with warm, humid boundary-layer air. Thus, strong wind shear helps thunderstorms sustain themselves for longer durations and to grow stronger.

A trigger refers to any process that forces an air parcel to rise through a cap of stable air and result in thunderstorm development. Triggers can be caused by frontal lifting, orographic effects, or surface heating.

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13.3: Atmospheric Instability and Thunderstorms New Page

Because the potential for thunderstorms to develop depends on atmospheric stability and layering, atmospheric soundings (e.g., Skew-T log-P) are used by meteorologists to help forecast storms. The soundings receive their data from the rawinsonde balloon launches, aircraft observations, dropsondes, satellites, or other meteorological data sources.

In atmospheric soundings, there are several labels indicating atmospheric stability. The labels are lifted condensation level (LCL), level of free convection (LFC), equilibrium level (EL), convective available potential energy (CAPE), and convective inhibition (CIN).

Lifted Condensation Level (LCL) was discussed in previous chapters. It is the altitude where the temperature cools to the dew point temperature, resulting in saturation and condensation. The LCL is the location where the cloud base of a thunderstorm develops.

Level of Free Convection (LFC) is the height at which the environmental temperature rate decreases faster than the air parcel's moist adiabatic lapse rate. This leads to atmospheric instability. In an atmospheric sounding, the LFC can be found where the moist adiabatic lapse rate of the rising parcel returns to the buoyant or warmer side of the environmental temperature.

Equilibrium Level (EL) is the height in the atmosphere where the temperature of the rising air parcel is the same as the temperature of its surroundings. The EL caps the atmospheric instability. This is where the moist adiabatic lapse rate returns to the negatively buoyant colder side of the environmental temperature. This is also where the anvil top of the thunderstorm is typically located.

Convective Inhibition (CIN) is the amount of energy that prevents a rising air parcel from reaching the level of free convection. On a thermodynamic diagram (Skew-T Log-P), it is the negative area between the environmental lapse rate and the air parcel lapse rate. CIN must be overcome for CAPE to be realized.

Convective Available Potential Energy (CAPE) is the amount of energy an air parcel would have if lifted vertically through the atmosphere over a particular distance. It is an indicator of atmospheric instability. On a thermodynamic diagram (Skew-T Log-P), CAPE can be found by the positive area between the environmental lapse rate and the air parcel lapse rate. It is an integrated measure of the total amount of buoyancy available to a rising air parcel.

CAPE can be used to estimate the maximum updraft velocity in thunderstorms.

$$\text{Maximum Updraft Velocity} = \sqrt{2 \cdot \text{CAPE}}$$

While the above equation gives a good estimate, the updraft velocity equation typically gives an unrealistically high value as it ignores many important processes. For example, dry air entrainment, liquid-water loading, and frictional drag are ignored. Observational studies find that the typical updraft velocity is roughly half of the maximum updraft velocity value.

$$\text{Likely Updraft Velocity} = \frac{\text{Maximum Updraft Velocity}}{2}$$

Thunderstorms are important features of Earth's atmospheric system. It is safe to say that there is always a thunderstorm occurring somewhere on Earth at all times. In many places, thunderstorms provide needed rain and are an important piece of the hydrological cycle. However, thunderstorms can also be hazardous and impacts will be discussed in the following chapter.

Chapter 14: Questions to Consider

1. Label the typical lifecycle of a single cell thunderstorm:
2. Describe the importance of updraft separation from the downdraft and precipitation portion of a thunderstorm.
- 3.
- 4.
5. With a CAPE of 1280 J/kg, calculate the likely updraft velocity.
6. Describe the importance of wind shear for thunderstorms.

Selected Practice Question Answers:

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CHAPTER OVERVIEW

14: Thunderstorm Hazards

Alison Nugent and David DeCou

Learning Objectives

By the end of this chapter, you should be able to:

- Identify the different types of precipitation (rain, snow, graupel, hail, etc.)
- Describe how hail forms and why it is difficult to predict
- Summarize how lightning occurs with respect to thunderstorm ice processes
- Discuss the various types of hazards from thunderstorms (gust fronts and outflow winds, precipitation, thunder and lightning, tornadoes)
- Identify radar hook echoes that suggest tornado hazards and discuss the process of warning the public about a tornado threat

[14.1: Introduction](#)

[14.2: Precipitation and Hail](#)

[14.3: Gust Fronts](#)

[14.4: Lightning](#)

[14.5: Tornadoes](#)

[14.6: Keeping Yourself Safe](#)

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14.1: Introduction



Lightning interacts with a tornado in Texas ([CC BY-SA 4.0](#)).

A thunderstorm is generally defined as any storm that has thunder and lightning. In fact, a climatological **thunderstorm day** is defined as any day on which thunder is audible from an observing station, and precipitation is not a required element. However, thunderstorms almost always produce precipitation along with the required thunder and lightning, and many other aspects that can be hazardous. Thunderstorms can produce damaging surface winds, hail, heavy rain, and even extreme phenomena like tornadoes. People are often most interested in the damaging aspects of severe weather, as these most directly impact human activity, and often produce beautiful and terrifying visual displays. The hazardous weather associated with thunderstorms will be the topic of this chapter.

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14.2: Precipitation and Hail

Thunderstorms consist of deep convective clouds that can produce large raindrops (2–8 mm diameter) within 5–10 km wide rain shafts that move horizontally across the ground with precipitation that last from 1–20 minutes over a fixed point. The rainfall rate may be heavy, as much as 10 to 1000 mm of rain per hour. The below photo depicts heavy rain shafts produced by a supercell thunderstorm in Montana.



A supercell thunderstorm in Montana ([CC BY 2.0](#)).

The cloud tops of thunderstorms extend far up in the troposphere. In the upper parts of the clouds, ice-phase processes occur. Throughout the cloud, different sizes of ice crystals and water droplets exist, and the heavier hydrometeors fall faster than the smaller ones, sometimes colliding and collecting each other. If these heavier ice particles fall through regions of supercooled droplets (liquid water droplets below freezing), they may grow through a process called **riming**. The liquid water droplets instantly freeze and adhere on contact to the outside surface of falling ice particles. This forms snow pellets called **graupel**, which are less than 5 mm in diameter. Fallen graupel is pictured below.



Graupel pellets on a car roof ([CC BY 3.0](#)).

Alternatively, smaller ice crystals that fall just below the 0°C level may partially melt and stick to other partially melted ice crystals through collision, causing them to grow through **aggregation**. These snow aggregates can grow as large as 1 cm in diameter. Snow aggregates and graupel can sometimes reach the ground still frozen or partially frozen, even in the summer, if they fall within cooler, saturated downdrafts. More frequently, these larger ice particles will melt completely into large raindrops before hitting the ground.

Hailstones are irregularly shaped balls of ice larger than 0.5 cm in diameter that are produced in a cumulonimbus cloud—particularly in intense, severe thunderstorms. Hail is formed when graupel or other particles act as embryos that grow through the accretion of supercooled liquid droplets in a cloud. One raindrop takes about one million cloud droplets to form, but a single hailstone may take as many as 10 billion cloud droplets to form. A golf ball-sized hailstone, as pictured below, must travel through the cloud for 5–10 minutes to be able to grow to this size. Strong, violent updrafts within the storm are able to carry smaller ice particles far above the freezing level where collisions with supercooled droplets allow them to grow into hailstones. Rotating

updrafts, such as those found in a supercell, are capable of sweeping hail horizontally through the storm, which may allow them to grow much larger. As these growing ice particles pass through areas of the cloud with high liquid water content, extra coatings of ice form around them. Larger hailstones may ascend very slowly in a strong updraft, almost “floating” in place as many supercooled water droplets continue to collide into them. When the hailstone is carried away from the updraft, or if it gets too heavy, it will fall because it is no longer supported by the rising air.

Just below the cloud, the hailstones will begin to melt in the warmer air. Small hail might completely melt before hitting the ground, but in the strong updrafts of severe thunderstorms, hailstones can grow large enough to reach the surface before melting. Notice in the figure below that the hailstone has concentric layers inside, similar to tree rings. The concentric layers alternate clear and opaque ice. This is related to the path that the hailstone takes when it passes through the cloud, as well as the liquid water content of the different areas the hailstone passes through. In areas with low liquid water content (dry growth regime), supercooled water droplets freeze onto the hailstone instantly, which produces the milky opaque ice that contains many air bubbles. In areas with high liquid water content (wet growth regime), supercooled water droplets collect very rapidly onto the hailstone. The transition from liquid to ice releases latent heat, which keeps the surface temperature of the hailstone at 0°C . As a result, supercooled droplets will form a coat of water around the stone rather than freezing instantly. This allows porous areas to be filled in as the water coat freezes slowly and air bubbles escape from the surface. This leaves a layer of clear ice around the hailstone.



Golf ball sized hail cut in half so the inside concentric layers can be viewed ([Public Domain](#)).

Large hail can cause severe damage to crops, vegetation, aircraft, roofs, windows, and create a traffic hazard if hailstones accumulate on a roadway. Damage can be greater if the strong winds in a thunderstorm cause hailstones to move horizontally as they fall. Larger hailstones have higher terminal fall velocities and the potential to reach speeds higher than $50\text{ m}\cdot\text{s}^{-1}$ (112 mph).

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14.3: Gust Fronts

Gust fronts were already discussed in the prior chapter, but we'll include a section here as well. Recall that mature thunderstorms contain both updrafts of warm, buoyant moist air as well as downdrafts of chilled, dry, dense air. Downdrafts are the result of a combination of evaporative cooling, precipitation drag, and dry air entrainment into the cloud. With the onset of precipitation in a mature thunderstorm, there is often a downward rush of cold air. As this cold air hits the surface, it spreads out along the surface in all directions. This surface boundary between the cold horizontally rushing air and the warmer air around it is called the **gust front**. Along this boundary, winds rapidly change both in direction and speed. The gust front is the leading edge of the outflowing cold air at the surface, and resembles a mini cold front to surface observers. If the gust front of an advancing thunderstorm were to pass by you, you would experience a sharp drop in temperature accompanied by a shift in wind direction and an increase in wind speeds, which may occasionally exceed 55 knots. The air is extremely turbulent along the leading edge of the gust front. As the cold air pushes warm, moist air upward along the boundary, a **shelf cloud** (or *arcus cloud*) may form, as shown in the below figure.

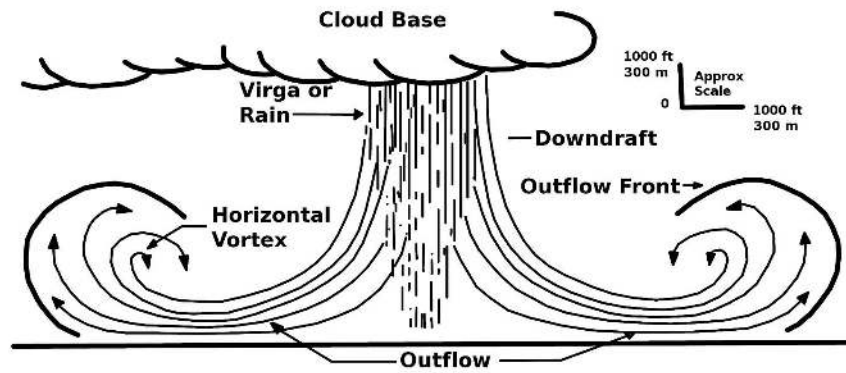


A shelf cloud, or arcus cloud, on the edge of a gust front ([CC BY-SA 3.0](#)).

When the atmosphere is conditionally unstable, the gust front may produce additional multicell thunderstorms along its leading edge as it forces warm, moist air upward. Each of these additional thunderstorms will each have their own gust front, which may merge together into a large gust front known as an **outflow boundary**.

Downbursts and Microbursts

Under a severe thunderstorm, the downdraft may become narrow, intense, and localised. As it reaches the surface, it may spread radially along the ground in a burst of wind—similar to water rushing from a faucet and hitting the bottom of the sink. This sort of downdraft is called a **downburst**. A short downburst, with winds of 4 km or less, is called a **microburst**. Despite being small spatially, a strong microburst can contain powerful winds up to 146 knots with the capacity to down trees and heavily damage weak structures and boats. Some damage caused by microbursts may have been incorrectly attributed to tornadoes. In addition, wind shear, turbulence, and gusts associated with microbursts have been responsible for several airplane crashes. Microbursts are often associated with severe thunderstorms but have also been known to occur with ordinary cell thunderstorms and even clouds that produce only scattered showers.



Cross-section of a microburst ([Public Domain](#)).

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14.4: Lightning

Lightning is caused by a discharge of electricity that typically occurs in mature thunderstorms. Lightning can strike within the same cloud, from one cloud to another, from a cloud to open air, or from cloud to ground. Only about 20% of lightning strikes are cloud-to-ground (CG) strikes, while the majority of strikes occur within the cloud (CC). As a lightning stroke travels through the atmosphere, it heats the surrounding air to a staggering $30,000^{\circ}\text{C}$, which is roughly five times hotter than the surface of the sun. This incredible heating causes a sudden, explosive expansion of the air, generating a shock wave in all directions from the flash. This booms outward in the form of a sound wave called **thunder**.

Due to the speed that light travels, we see lightning instantly as it happens. Thunder travels much more slowly at about $330\text{ m}\cdot\text{s}^{-1}$ so it takes much longer to reach our ears. You can estimate how distant a lightning stroke is by counting the seconds after you see the flash until you hear the thunder. It takes sound about 5 seconds to travel one mile. If you hear thunder roughly 10 seconds after you see the lightning flash, then the strike occurred about 2 miles away. However, if the lightning is too far, you may not hear any thunder, because sound waves attenuate and bend as they travel through the atmosphere, especially through layers of differing air temperature.



Lightning during a desert storm ([CC 0](#)).

Why do we have lightning? During ordinary fair weather conditions, the earth's surface has negative charge, while the upper atmosphere is positively charged. Lightning requires differing regions of opposite charge inside a cumulonimbus cloud. There are several theories as to how this separation of charge may come about.

One theory is that clouds can become electrified when graupel and hailstones fall through areas of supercooled droplets, causing the droplets to freeze and release latent heat. This release of latent heat keeps the hailstone surface warmer than the surrounding ice particles. A net transfer of positive ions (molecules that carry a charge) occurs from the warmer hailstone to the colder surrounding ice crystals when they come in contact. Therefore, the larger, warmer hailstone becomes negatively charged and the smaller, cooler ice crystal becomes positively charged. This also happens when supercooled liquid droplets freeze when they come into contact with a warmer hailstone and little fragments of positively charged ice break off. These tiny positively-charged particles are easily swept to the upper regions of the cloud through updrafts. The larger, heavier, negatively-charged hailstones tend to remain suspended at the same level in the updraft or fall to the lower parts of the cloud. This causes the colder upper regions of the cloud to have a net positive charge, while the middle level of the cloud has negative charge. The lowest portion of the cloud generally has a mixture of negative and positive charges, with a positively charged region occasionally present in precipitation bands near the melting level.

In short, thunderstorms must have ice phase processes to produce thunder and lightning—lightning is produced from a discharge of electricity and thunder is simply produced as a result of the lightning.

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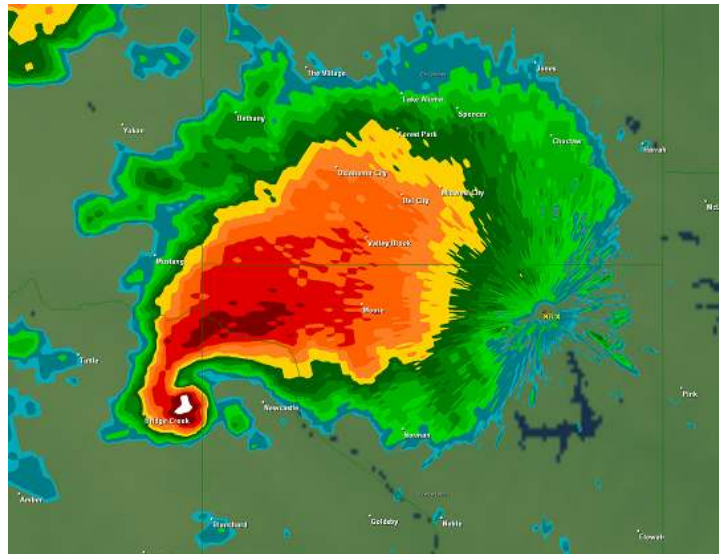
14.5: Tornadoes

Recall from the prior chapter that supercell storms can produce tornadoes. A supercell is a severe thunderstorm with rotation. Typically the rotation of a supercell thunderstorm comes from directional wind shear in the environment. Occasionally a supercell storm produces a tornado, a region of rapidly rotating air in cyclostrophic balance. Recall from Chapter 10 that cyclostrophic balance is a balance between the pressure gradient force and the centrifugal force. As the rotating air motion in a supercell is vertically stretched, it narrows and speeds up. To picture this, imagine a figure skater pulling their arms inward as they spin. Their diameter narrows and their rotation speed increases. The same physics—conservation of angular momentum—occurs as a tornado forms.



An EF 4 Tornado in Marquette, Kansas ([Public Domain](#)).

Tornadoes are difficult to forecast well in advance. Typically a Tornado Watch will be issued by a Storm Prediction Center when all of the ingredients for a tornado are present in the atmosphere. These ingredients are the same as for a severe thunderstorm: warm moist air; instability (quantified by large values of CAPE); potential for a trigger; and, of course, directional wind shear to provide rotation. A Tornado Warning is issued when a tornado is imminent or occurring. For example, one may be issued when a funnel cloud is spotted or when a “**hook echo**” is seen on radar. The issue of forecasting tornadoes is called “nowcasting”. Although very little advance notice can be given to the public, forecasters do their best to provide as much notice as possible.



Hook echo in a radar image (Public Domain).

The above image shows a hook echo from radar. The key feature is the bottom left region of the storm where a region of rain can be seen rapping around a region of dry inflow air. Right at the cusp of the hook is where a tornado would form if it were occurring. The strong rotation in a tornado creates this radar signature.

Tornado strength is categorized after they occur. A scale called the **Enhanced-Fujita scale** is used to classify tornado strength based on the amount of damage that was caused. While the wind speed for each step in the EF scale is given, it is not measured. The wind speed is determined afterwards by the scale of damage and the winds required to cause the damage.

Enhanced Fujita Scale

Scale	Wind Speed (3 sec. gust, mph)	Damage
EF0	65-85	Minor
EF1	86-110	Moderate
EF2	111-135	Considerable
EF3	136-165	Severe
EF4	166-200	Extreme
EF5	>200	Massive/Incredible

The central US is the world's hot-spot for tornados, however, tornados and tornado-like storm features can occur elsewhere. A dust devil and a waterspout are tornado-like features that are associated with cumuliiform and sometimes cumulonimbus clouds. However, dust devils and waterspouts typically are not associated with a supercell. Waterspouts look like tornados but form over water surfaces. Waterspouts are often narrow, relatively short lived, and formed by the same physics as a tornado—a rotating updraft with condensation in the funnel cloud. Below is an image of a waterspout observed in Hawaii off the coast of Oahu.



Waterspout spotted off O'ahu's southern shore ([Taken on May 2, 2011, by Staff Sgt. Mike Meares](#)).

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14.6: Keeping Yourself Safe

From reading this storm hazards chapter, hopefully you've learned that severe storms and storm hazards should be taken seriously. Part of the issue with severe weather forecasting is that the impacts are often extremely localized. It is difficult to know if you will be in the exact location that hail will fall or a tornado will pass. Oftentimes, you'll simply be aware that the possibility is there for storm impacts. Therefore, it is important to take precautions, regardless of whether you are in the right place at the wrong time.

For example, here are a few simple steps you can follow:

- “Turn around, don’t drown” is the saying from the National Weather Service. If you see flowing water over a roadway, don’t drive over it!
- If there is a possibility for thunderstorms, don’t go hiking or do outdoor activities near streams or rivers. Flash flooding is a possibility.
- Similarly, if there is a possibility for thunderstorms, don’t put yourself in an exposed outdoor location. As soon as you hear thunder, a thunderstorm is close enough to be a hazard. Get yourself off of that open sports field, beach, or stadium, and into a shelter until it passes.
- If a tornado is approaching, the safest place is inside the interior room in your house (unless you have a tornado cellar). Get in your closet or your bathtub with a mattress or something to cover you.

Depending on where you live in the world, some hazards are more likely than others. Still, it is important to think in advance about what you would do if you were in a severe storm situation. Oftentimes there isn’t a lot of time to react.

Chapter 15: Questions to Consider

1. Why does hail have concentric layers that alternate clear and opaque ice?
- 2.
- 3.
4. *FLASH* You saw a lightning stroke! 15 seconds go by... *BOOM* You heard thunder! How far away was the lightning stroke?
5. Where is the tornado?
6. Bob just finished reading this chapter and wants to practice sheltering in the event of a tornado warning. Can you help him find the safest place to shelter in his house?

Selected Practice

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CHAPTER OVERVIEW

15: Atmospheric Optics

Learning Objectives

By the end of this chapter, you should be able to:

- Explain why the sky is blue and clouds are white
- Discuss why lots of rainbows are seen on Hawai'i
- Describe how one can observe a rainbow
- Determine from visibility whether objects are close or far away

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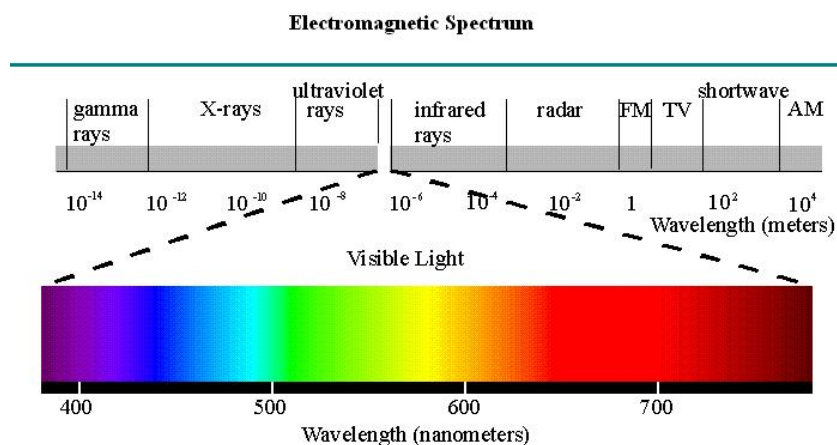
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15.1: Introduction



Rainbow over the Ala Wai Canal (Photo by Shintaro Russell).

The color of our world is defined by the interaction of objects with light. In fact, the reason we can see anything at all is due to the interactions of objects with light. Light can be thought of as both a particle and a wave. In this chapter, we'll consider light only as a wave defined by some wavelength on the electromagnetic spectrum. Recall the electromagnetic energy spectrum from Chapter 2 with the visible portion near the center between ultraviolet and infrared:



The electromagnetic energy spectrum showing the wavelength of colors ([CC BY-NC-ND 2.0](https://creativecommons.org/licenses/by-nc-nd/2.0/)).

Notice that blue and purple colors have a relatively short wavelength (~ 400 nm) and orange and red have a relatively long wavelength (~ 700 nm). While slight, this difference makes a large difference in the interaction of light with particles in our atmosphere.

Before we begin, let's define a few important facts that will help us to understand the following material.

1. White light contains **all** colors. Light from the sun is considered white light.
2. The color an object appears to be is due to reflection of that color toward your eye. For example, a green leaf looks green because it absorbs red and blue light, but reflects the wavelength of light that is green.

Based on this simple description, we can say that the sky appears to be blue because the molecules in the atmosphere reflect blue light. But why do they reflect blue light more than red light? This has to do with the relative sizes of the molecules and the wavelengths of light interacting with them.

In general, there are three different laws regarding wavelength and interaction with particles. We'll define λ as the wavelength of light, and d as the diameter of the molecule or particle.

When the wavelength of light is much much larger than the diameter of the molecule or particle, Rayleigh scattering is dominant.

$$\text{Rayleigh Scattering } \lambda \gg d$$

When the wavelength of light is within an order or two of magnitude as the diameter of the molecule or particle, Mie scattering is dominant.

$$\text{Mie Scattering } \lambda \sim d$$

Finally, when the wavelength of light is much much smaller than the diameter of the molecule or particle, Geometric Optics is dominant.

$$\text{Geometric Optics } \lambda \ll d$$

Rayleigh Scattering

Rayleigh scattering is the reason the sky is blue. Rayleigh scattering occurs when the wavelength of visible light is much much larger than the particles it is interacting with. Such is the case with the tiny molecules in Earth's atmosphere. The amount of scattering, " S ", is proportional to:

$$S \sim \frac{1}{\lambda^4}$$

Because this relationship is dependent on wavelength (λ), shorter wavelengths will be scattered more. Purple and blue light have a shorter wavelength than orange and red light and are therefore scattered more, accounting for the blue color of Earth's atmosphere.

Mie Scattering

Mie scattering is the reason clouds are white. Mie scattering occurs when the wavelength of visible light is approximately the same as the particle or droplet diameter (within a factor of 10 or 100). When this is true, all wavelengths of visible light are scattered roughly equally. In this case, the color illuminating is the same color reflecting. In the case of clouds, they appear white because white light is coming from the sun and illuminating them.

Geometric Optics

Finally, geometric optics is responsible for rainbows. A beam of light is called a ray. The tracing of the ray path through a raindrop is called geometric optics and helps us to understand how and why rainbows form. When light comes in contact with a density interface, for example air and water, it can reflect or refract. Refraction is responsible for everyday conundrums like the following image.



Refraction of light at an air-water interface due to the difference in density between the two media ([CC BY-NC 2.0](#)).

Refraction occurs when rays bend at a density interface. Reflection occurs when rays bounce back from an object. These two phenomenon are responsible for why we can see atmospheric optical phenomena such as rainbows, halos, and sundogs. Here we will focus only on rainbows, the geometric optics occurring in liquid rain drops.

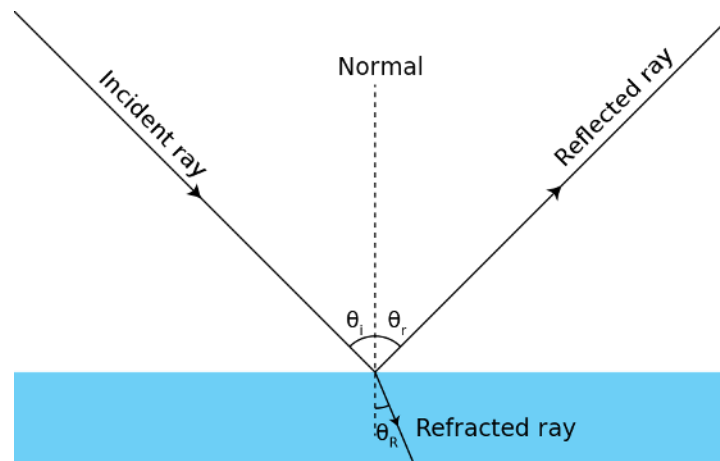
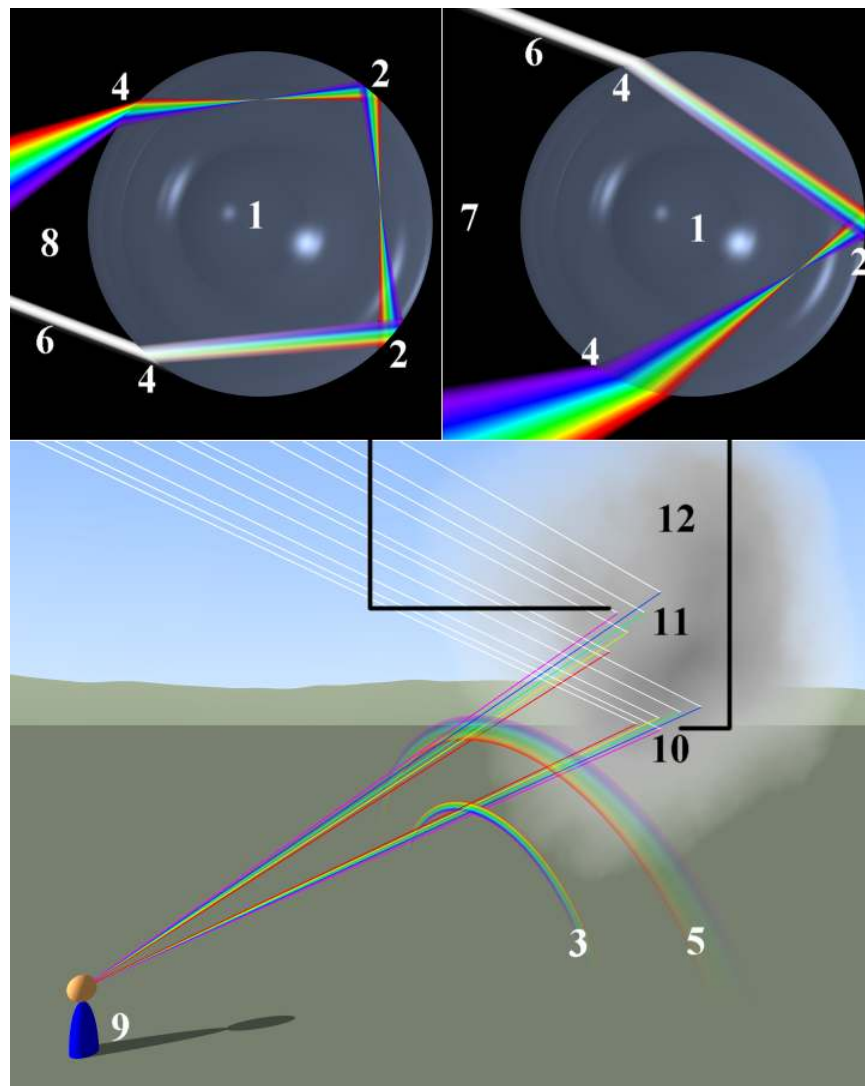


Diagram showing angle of incidence, reflection, and refraction of a ray of light ([Public Domain](#)).

As a light ray makes contact with an interface between two media with different densities like air and water, some of the incoming (incident) light will either refract from the air into the water, reflect back into the air, or absorb into heat. In the case of light interaction with a rain drop, we have a combination of refraction and reflection. A single rainbow has two refractions and one reflection, and the second rainbow seen in a double rainbow has two refractions and two reflections.

The top right image of the following figure shows the interaction of light with a rain drop responsible for a single rainbow. Light comes in contact with a rain drop, refracts as it enters the rain drop, reflects off the back side of the rain drop, and refracts again as it exits the rain drop. The two refractions result in a splitting of the colors of light, like a prism. This is because refraction of light is dependent on wavelength; shorter wavelengths refract more than longer wavelengths.



The formation of a rainbow with light propagation. Legend: 1.) Spherical droplet 2.) Places where internal reflection of the light occurs 3.) Primary rainbow 4.) Places where refraction of the light occurs 5.) Secondary rainbow 6.) Incoming beams of white light 7.) Path of light contributing to primary rainbow 8.) Path of light contributing to secondary rainbow 9.) Observer 10.) Region forming the primary rainbow 11.) Region forming the secondary rainbow 12.) Zone in the atmosphere holding countless tiny spherical droplets ([CC BY-SA 3.0](https://creativecommons.org/licenses/by-sa/3.0/)).

Put concisely, a rainbow is an atmospheric phenomenon displaying a spectrum of light as a result of refraction of light in water droplets. This is possible because the wavelength of light is much much smaller than the size of the rain drop.

Primary rainbows have red on the outside of the circle and purple on the inside. There is always a 42° angle between the incident light, the primary rainbow, and the observer. This constant angle can be seen in the above figure. At a given time, rainbows appear different to different observers because of different interaction between incoming light rays and water droplets. For instance, one observer may see a well-defined rainbow in an area of large rain drops and bright sun. Another observer in a half kilometer away with smaller rain drops may see a partial rainbow.

Secondary rainbows are outer rainbows with red on the inside of the circle and purple on the outside because of the double reflection. Secondary rainbows occur at a 52° angle, hence secondary rainbows are higher than primary rainbows. They're not as bright as primary rainbows, also because of the double reflection. Triple rainbows are also possible and have been observed, but are much less frequent.



Double rainbow in a stormy sky. Notice how the colors are flipped in the two rainbows, and how the secondary rainbow is not as bright as the primary rainbow (Photo by Shintaro Russell).

How to see Rainbows

The best chance of seeing a rainbow is when there is rain and sun low in the sky at the same time. An observer may see a rainbow if the sun is at the observer's back and they are facing towards a region that is raining. As the sun lowers in the sky, the height of the rainbow will increase because of the constant 42° angle. Rainbows cannot be seen on the ground at noon because a 42° angle is not possible. Rainbows appear as half arcs only because the ground typically blocks the other half. Rainbows are frequent in Hawai'i because of the abundant sunlight and frequent rain.

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15.2: Visibility

Visibility is the final topic on the interaction of light with particles in Earth's atmosphere covered in this chapter. In general, the topic of visibility is simply a question of what can be seen and how far away it is. When someone says that the visibility today is 10 miles, it means that one can see objects 10 miles away.

In the Arctic, where the air is extremely dry and clean, visibility of 100 miles (161 km) is common. On the other end, when you cannot see objects 100 meters (~330 ft) away this is referred to as zero visibility. As you may imagine, visibility is extremely important for pilots and drivers. Fog in the atmosphere provides the worst conditions for visibility.

Look at the following image of the Great Smoky Mountains. The trees in the front of the image are nearly clear, but with each additional ridge, the mountainside looks whiter and whiter. We assume that each ridge has the same type of foliage as the first. This tells us that objects further away are brighter.

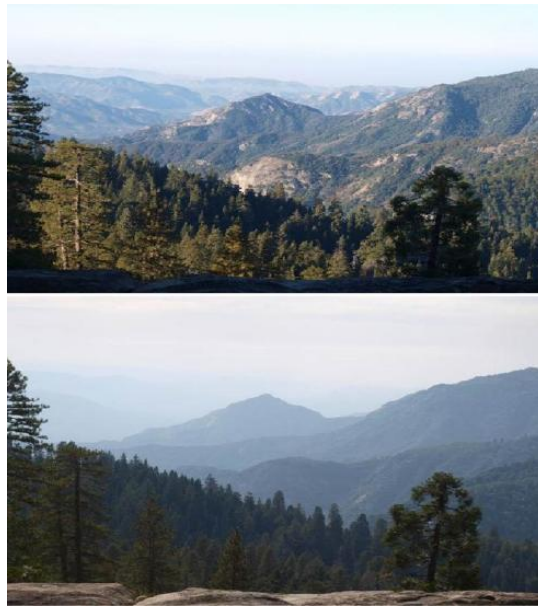


A photo from the Great Smoky Mountains National Park illustrating that objects far away appear brighter ([CC BY 2.0](#)).

The reason for the change in brightness is particulate matter in the atmosphere. This can come in the form of aerosols or hydrometeors (cloud droplets, rain drops, snow flakes etc.). In the case of the mountain image above, the Great Smoky Mountains are known for emitting plentiful volatile organic compounds (VOCs) that act as cloud condensation nuclei (CCN) and produce clouds near ground level (i.e., low hanging fog). Both the aerosols and the hydrometeors reduce visibility.

Many believe that a lack of visibility is caused by objects blocking your view. However, if that were the case, the background would appear darker. Instead, a lack of visibility is caused by objects scattering light towards your eye. With greater distances, your eye has to look through a larger portion of the atmosphere and, therefore, more opportunity for light to scatter towards you. The scattered light is white because *Mie scattering* is taking place.

Here is another image showing the same scene on a clear and polluted day from the Sequoia and Kings Canyon National Park.



A clear (top) and polluted (bottom) day at Sequoia & Kings Canyon National Park ([Public Domain](#)).

This national park often has issues with pollution from the central valley in California. On polluted days, more aerosols and particulate matter are in the atmosphere, so more particles are available to scatter light toward your eye. Scene details are lost and the mountains in the distance can't be seen at all. The visible range is shorter on the bottom image as compared to the top.

Chapter 22: Questions to Consider

- 1.
- 2.
3. Why is it not possible to see a rainbow on the ground at noon?
4. Drag the Sun, Rain Cloud, Rainbow, and Bob to the correct positions in the situation that Bob sees a rainbow when looking East, assuming East is the right side of the picture:
5. What causes low visibility? What type of scattering is this an example of?

Selected Practice Question Answers:

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