# An Introduction to Weather and Climate

**Neel Desai and Alicia Mullens**

De Anza College

Geosciences

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This text was compiled on 10/23/2026

# Program Page

De Anza College sits in the heart of Silicon Valley, where innovation and opportunity are part of daily life, and also has one of the highest costs of living in the country. For our students, that makes every dollar count, and the price of a textbook can be the difference between staying enrolled or dropping a class. Our Zero Textbook Cost (ZTC) program closes that gap by giving students full access to high-quality, faculty-created course materials at no cost from the first day of class.

De Anza is proud of its diverse student community, and we hold our ZTC materials to the same standard of excellence that defines the college. Faculty across departments write, adapt, and review these resources themselves, keeping them accurate, current, and grounded in the perspectives of the students who use them, with accessibility and inclusive design built in from the start.

This program reflects the time and care our faculty invest in their students. If you have questions, feedback, or come across any issues with a ZTC resource, contact De Anza College's ZTC Grants Coordinator, Shagun Kaur.

Learn more about De Anza ZTC Program [www.deanza.edu/academic-services/oer/]

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## Detailed Licensing



## Detailed Licensing



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# Accessibility Statement

## Accessibility

De Anza College is dedicated to creating learning materials that are accessible and benefit all students. Our materials follow the Web Content Accessibility Guidelines (WCAG) 2.1 Level AA. Our authors and contracted accessibility specialists checked all resources using this ASCCC OERI checklist, the accessibility tool in LibreTexts, WAVE, JAWS, and other more detailed checks for complex materials such as interactive components.

We selected LibreTexts as our preferred platform because of its commitment to accessibility. In addition to the platform's inherent accessibility, we collaborated with the LibreTexts team to fix any accessibility issues we found.

## Our Accessibility Review Process

Accessibility was built into the resource as it was written. This resource went through the following stages of review:

1. **Template check.** The project coordinator and instructional designers checked the first chapter for accessibility as a model to ensure later chapters meet accessibility standards.
2. **Progress checks.** The manuscript was checked at approximately 40%, 60%, and 80% completion to catch and correct accessibility issues in production.
3. **Specialist review.** Once the manuscript was complete, an accessibility expert conducted an accessibility check.
4. **Remediation.** The author and the project manager who conducted the earlier progress checks worked together to remediate the issues identified by the accessibility expert.
5. **Final check.** The resource underwent another full check by the accessibility expert before the book was deemed ready for publication.

This process combines ongoing checks during writing with a final specialist review, so issues are caught early and confirmed fixed before publication.

## Report an Accessibility Problem

We are always looking for ways to make our resources more accessible. If you run into a barrier while using this resource, please let us know so we can look into it. When you contact us, it helps to include:

- The name of the resource
- The location of the problem (a web address or page/chapter description)
- A description of the problem you encountered
- The device, browser, and any assistive technology you were using (e.g., Windows 11, Google Chrome, JAWS screen reader)

**Contact us:** De Anza ZTC team

*This statement was last updated August 2026*

This resource was project-managed by **Shagun Kaur**, overseeing scope, budget, and schedule; stakeholder coordination via monthly meetings; training and platform support (LibreTexts, troubleshooting, technical support); QA/accessibility check scheduling; and licensing and copyright governance.

## A Note on Known Limitations

No accessibility check is perfect. We've made a genuine, sustained effort to catch and fix issues throughout the writing and review process described above, but some barriers may remain, especially in complex or interactive content. We'd rather be upfront about that than claim a standard of perfection we can't guarantee.

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# About the Textbook

This **textbook** is part of the **De Anza College Zero Textbook Cost (ZTC) Pathways** initiative, supporting both students and instructors with free, adaptable Open Educational Resources (OER).

Few, if any open educational resources (OER) exist for entry-level meteorology students, including both majors and non-majors. This book attempts to close that gap with a resource specifically designed for introductory-level qualitative survey courses. The bulk of the book (Chapters 1-11) has been adapted from the University of Hawai'i's ER textbook, "Atmospheric Processes and Phenomena," by Alison Nugent, David DeCou, Shintaro Russell, and Christina Karamperidou, while Chapter 12 was adapted from Andreas Schmittner, from Oregon State University's OER textbook, "Introduction to Climate Science." Except for Chapter 3, the primary modifications to Nugent's book were simplifying technical and quantitative descriptions into more qualitative descriptions appropriate for non-majors and beginning students with little background in mathematics. Chapter 3 was a near-complete rewrite, changing the focus of Nugent's book from Thermodynamics to Air Temperature, applications of air temperature, and seasonal factors that affect air temperature. Chapter 12 was primarily modified to meet accessibility guidelines, with the bulk of Schmittner's work left unchanged. The final result is a comprehensive introduction to the fields of Meteorology and Climate Science for non-majors, entry-level students, and interested people in the general public.

Without these foundational texts, a lot more work would have been required to complete this project.

## Acknowledgments

The authors would like to thank their respective spouses for their infinite patience and support during the writing of this textbook. A special thanks also to the ZTC coordinators and the OER team at De Anza College for their guidance during the publication process.

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# About the Authors

*Information about the authors*

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|---|---|
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|  | **Neel Desai** is a Meteorology instructor at De Anza College in Cupertino, CA, and at San Jose State University in San Jose, CA. He is passionate about Meteorology and Climate Science, and strives to provide an enjoyable in-class learning experience for his students. He believes in a visual and hands-on approach to learning, incorporating multiple videos/in-class activities to keep students engaged. He has experience teaching students at multiple levels, including high school-to-university bridge programs at the University of Michigan, community college at De Anza, and the university level, including graduate students at SJSU. Neel received his Bachelor's degree in Mechanical Engineering from the University of Mumbai, India, a Master's degree in Aerospace Engineering from the University of Michigan, Ann Arbor, and a doctorate in Atmospheric Sciences from Michigan Technological University in Houghton, MI. His research interests include cloud physics, turbulence, and remote sensing. Outside of teaching and research, Neel enjoys traveling, hiking, and reading sci-fi novels. |

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# 1: EARTH'S ATMOSPHERE

## 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 about the vast volume of air around us that allows us to do so. The atmosphere brings us the oxygen that fills our lungs; it brings us the beautiful blue 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.

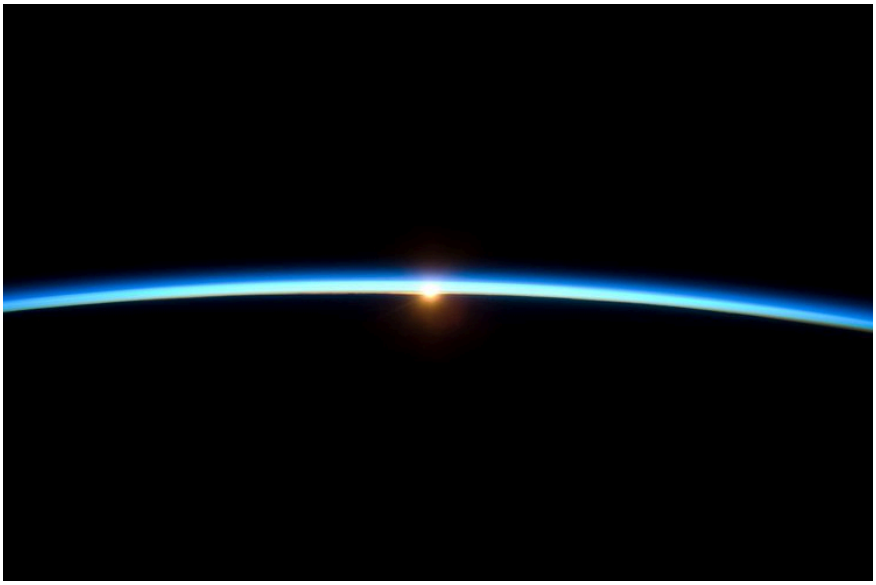


Figure 1.1: View from outer space of the sun rising over Earth, illuminating the atmosphere in a ring of blue. (Public Domain; NASA, via Flickr).

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 appears as a 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.



Figure 1.2: View of the moon from space over Earth's blue atmosphere (Public Domain; NASA, via Wikimedia Commons).

The goal of this chapter is to orient readers to the basics of our atmosphere.

## Learning Objectives

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

- Name the gases that comprise the Earth's atmosphere
- Describe the vertical structure of Earth's atmosphere with respect to pressure and temperature
- Discuss the difference between weather and climate


1.1: Overview of Earth's Atmosphere

1.2: Equation of State — Ideal Gas Law

1.3: Layers of the Atmosphere

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

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

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# 1.1: Overview of Earth's Atmosphere

## Energy in the Atmosphere

Everything that happens on Earth is, in some way, caused by radiation from the sun, 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 the 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

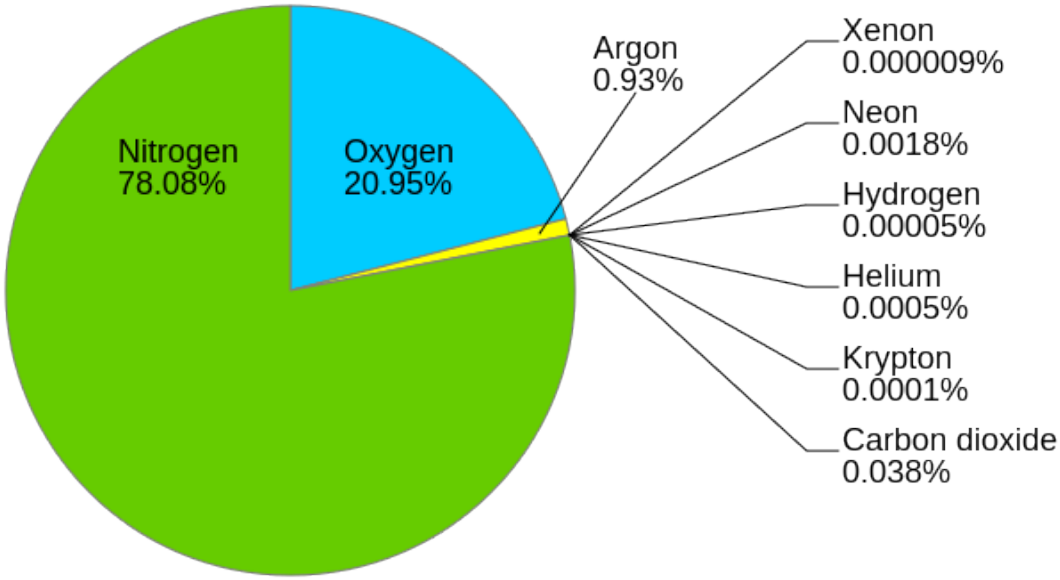



Figure 1.1.1: Pie chart showing the gases that make up Earth's atmosphere. A breakdown is provided in the text. (CC BY-SA 3.0; Life of Riley, via Wikimedia Commons). Alternative Description of Image.

Earth's atmosphere is primarily composed of nitrogen ($N_2$) and oxygen ($O_2$), with smaller quantities of other gases, as shown in the figure above. Nitrogen makes up around 78%, oxygen makes up around 21%, and Argon makes up almost 1% of the total dry air volume in the atmosphere. Other gases (such as water vapor, the gaseous form of water) 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 and releases carbon dioxide ( $CO_2$). When plants photosynthesize, they take in carbon dioxide and release oxygen. When water ($H_2O$) evaporates from the ocean, additional water vapor (gas) is added to the atmosphere. When a volcano erupts, sulfur dioxide ($SO_2$) is added to the atmosphere (Figure 1.1.2).

## Water Vapor

Water vapor concentrations vary significantly from location to location and from time to time. In warm, moist tropical regions, water vapor can make up nearly 4% of the atmosphere, while in the 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 by which water vapor changes phase to liquid is called *condensation*. When liquid water becomes a gas, it is called *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 significant 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-radiates a portion of the Earth's outgoing radiation, helping our planet 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.

# 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 gravity, which pulls the atmosphere downward. 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 affects 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 not weigh 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 of Units (SI). 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 a 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 Earth's surface and decreases with height because there are more molecules held tightly together by gravity at the surface than above. The standard unit for density is $kg \cdot m^{-3}$.

**Pro Tip:** Density is often denoted by the Greek lowercase letter $\rho$ (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.

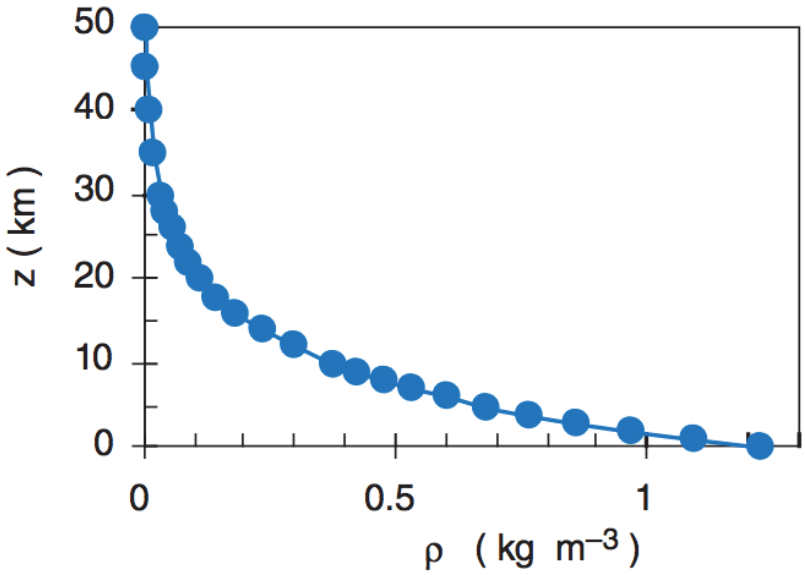



Figure 1.1.2: Density ($\rho$, kg·m$^{-3}$) vs. height ($z$, km) in the atmosphere (CC-BY-NC-SA 4.0; Roland Stull via Practical Meteorology). Alternative Text Description.

Figure 1.1.2 shows the distribution of density ($\rho$) with altitude. Atmospheric surface density is typically around 1.2 kg·m$^{-3}$. We call this value $\rho_0$ because it is the initial value at the surface. As you rise in the atmosphere, density decreases exponentially, with a steep drop at first, followed by progressively smaller decreases as you go higher. Table 1.1.1 is a table of the data from Figure 1.1.2.

*Table 1.1.1. Approximate density vs height.*

| z (km) | ρ (kg·m⁻³) |
|---|---|
| 0 | 1.20 |
| 2 | 1.00 |
| 4 | 0.82 |
| 6 | 0.66 |
| 8 | 0.53 |
| 10 | 0.41 |
| 12 | 0.31 |
| 14 | 0.23 |
| 16 | 0.17 |
| 18 | 0.13 |
| 20 | 0.10 |
| 22 | 0.075 |
| 24 | 0.055 |
| 26 | 0.040 |
| 28 | 0.030 |
| 30 | 0.022 |
| 35 | 0.012 |
| 40 | 0.006 |
| 45 | 0.003 |

| z (km) | ρ (kg·m⁻³) |
|---|---|
| 50 | 0.0015 |

## 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 billion 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.

> **Pro Tip:** You will generally find air pressure expressed in units of millibars (mb) or hectopascals (hPa) on weather maps. These two units are equivalent. In the aviation field, Inches of mercury (inHg) are also commonly used. In Engineering, the SI unit Pascal (Pa) is commonly used. At sea level, the global average for atmospheric pressure is:
>
> 101.325 kPa = 1013.25 mb = 1013.25 hPa = 29.92 in. Hg. = 1 atm (atmosphere) = 101325 Pa.
>
> It's important to know that air pressure is commonly measured in millibars/hectopascals in meteorology.

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 Earth's surface and fewer above, air pressure is greatest at the surface. It decreases with height nearly exponentially, analogous to air density.

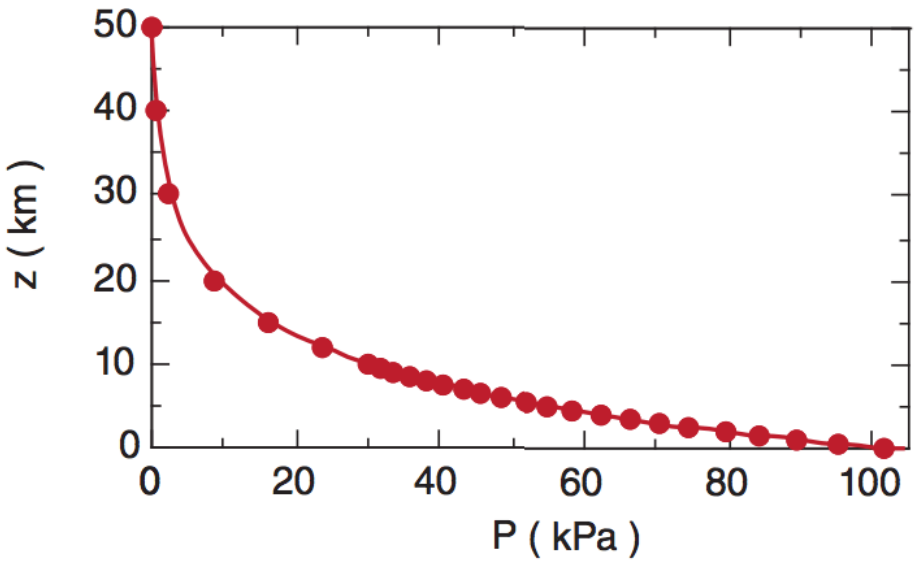



Figure 1.1.3: Height (z, km) vs. pressure (P, kPa) in the atmosphere. (CC-BY-NC-SA 4.0; Roland Stull, via Practical Meteorology). Alternative Text Description

Figure 1.1.3 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. Like density, air pressure decreases exponentially with height. Data for Figure 1.1.3 is provided in Table 1.1.2.

*Table 1.1.2: Air Pressure vs. Height*

| z (km) | P (kPa) |
|---|---|
| 0 | 101 |
| 1 | 90 |

| z (km) | P (kPa) |
|---|---|
| 2 | 80 |
| 3 | 70 |
| 4 | 62 |
| 5 | 55 |
| 6 | 48 |
| 7 | 42 |
| 8 | 37 |
| 9 | 33 |
| 10 | 30 |
| 12 | 24 |
| 15 | 18 |
| 20 | 10 |
| 30 | 4 |
| 40 | 1.5 |
| 50 | 0.5 |

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.

## 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 one you should be very familiar with: gravity.

In the atmosphere, when the vertical pressure gradient force balances gravity, the system is in **hydrostatic balance**. The word hydro means water or fluid, and static means stationary, so that the name can be interpreted as a stationary fluid balance. This balance holds in most atmospheric situations.

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# 1.2: Equation of State — Ideal Gas Law

## Equation of State

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, $\rho$) 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 behave ideally, meaning there are few intermolecular forces and that the molecules are small compared to the space between them. In almost all cases, atmospheric gases can be treated as ideal gases. Again, this is another example of a simplifying assumption we use to approximate relationships in the atmosphere.

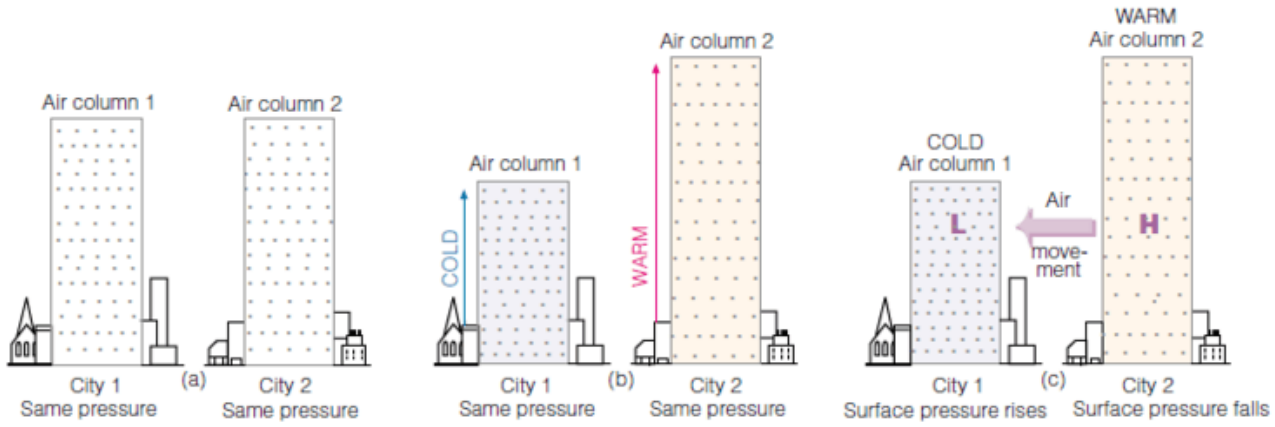



Figure 1.2.1: Air pressure differences associated with air temperature in a column (CC-BY-SA 4.0; Milya Dwi Lestari, via Wikimedia Commons).

Figure 1.2.1 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 at the surface is the same in both cities because it is determined by the number of air molecules above it.

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 faster, and the column's volume is larger. It takes less cold air to achieve the same surface pressure as a larger volume of warm air. 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, leading to air movement visible 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 City 1. Locally, this creates a higher pressure aloft over the warmer City 2. Because of the difference in pressure aloft, airflow is created that flows from the higher to the lower pressure. This airflow caused by a localized difference in air pressure is what we call wind.

Note

The force arising from 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 a later chapter.

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.

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# 1.3: Layers of the Atmosphere

## Vertical Profile of the Atmosphere

In the previous sections, we discussed how both air pressure and density decrease with height because most of the atmosphere is held close to 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 layers in several 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.

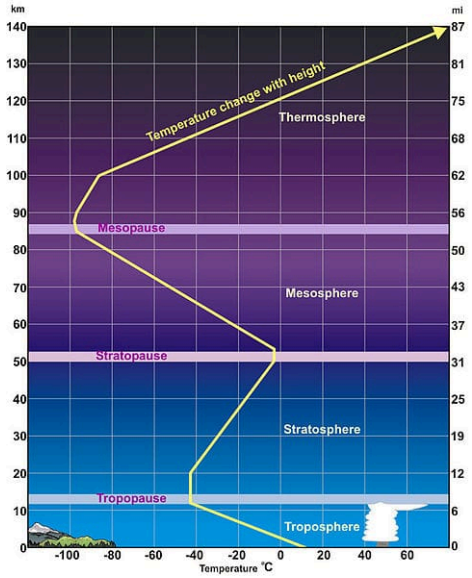



Figure 1.3.1: The vertical temperature structure of Earth's atmosphere in kilometers (left axis) and miles (right axis). The layers are labeled as "spheres" and the boundaries between the layers are labeled as "pauses". For example, the troposphere is the lowest layer, and the tropopause is the boundary between the troposphere and the stratosphere. The yellow line shows temperature in °C, and the cloud indicates where most of the weather occurs. (Public Domain; NOAA, via Wikimedia Commons)

As illustrated in Figure 1.3.1, 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. The layers of the atmosphere are described in the following sections.

## 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 the weather we experience on Earth occurs 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 with 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 proximity to the Earth's surface. Airflow closest to the surface slows due to friction, which can be amplified by

terrain or vegetation. As a result, the boundary layer experiences the most significant 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. Near the top of the atmospheric boundary layer is a section where temperature briefly increases with height, called an **inversion**. Inversions act like a lid, preventing air from above from mixing into the boundary layer. This can have profound implications on days when pollution is particularly problematic.

## Stratosphere

In the *stratosphere*, the air temperature increases with height, causing a temperature inversion. This inversion layer tends to prevent tropospheric air from rising and 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*, air temperature decreases with height, and the air is extremely thin and low-density. The average atmospheric pressure in this layer is about 1 mb, which means that about 99.9% of all air molecules are below this level, and only about one thousandth of the atmospheric mass is above it. 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 temperatures decrease with height in the mesosphere is that there is not much ozone to absorb UV radiation. As a result, 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, around -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 significant increase in temperature. Molecules can travel for entire kilometers without colliding with another molecule. This is the layer where auroras form due to interactions between charged solar particles and air molecules.

At the top of the thermosphere, 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.

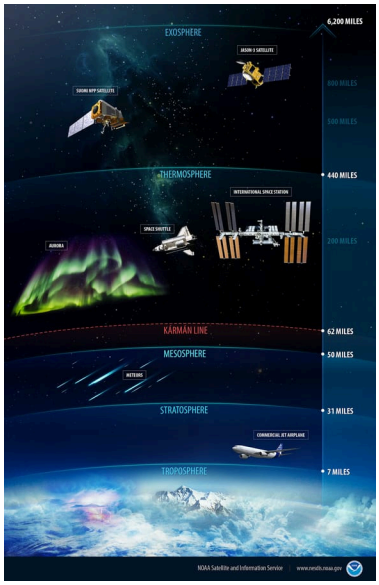



Figure 1.3.2: The relative heights of different phenomena in the various vertical layers of the atmosphere. The Kármán Line at 62 miles (100 km) designates the boundary between Earth's atmosphere and outer space. Alternative text in the paragraph below. (Public Domain; NOAA Satellite and Information Service via Flickr).

Figure 1.3.2 is an illustration showing the layers of Earth's atmosphere with examples of objects and phenomena found in each. From lowest to highest: the troposphere contains clouds and a jet airplane; the stratosphere contains a commercial jet airplane; the mesosphere shows meteors; the thermosphere shows auroras, a space shuttle, and the International Space Station; and the exosphere shows satellites. The Troposhere extends to an approximate altitude of 7 miles, while the Stratosphere extends to an approximate altitude of 31 miles. The Mesosphere extends to an approximate altitude of 50 miles while the Thermosphere extends to an altitude of 440 miles, and the Exosphere to an altitude above 6,200 miles. The Kármán line, commonly considered the boundary between the atmosphere and outer space is marked at 62 miles.

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# 1.4: Weather and Climate- What's the Difference?

## Difference Between Weather and Climate

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 constantly changing. 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 precisely 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 fog or precipitation.
- **Wind** — the horizontal flow of air, caused by local differences in air pressure.

In later chapters, we will discuss each of these weather variables and phenomena in more detail.

Meanwhile, **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 that 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 California, it isn't likely that you would need to buy a heavy winter jacket, while you will need to buy one in Minnesota or North Dakota.

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# 1.5: Chapter 1 Reference Guide- Coordinate Systems, Units, Terminology

## Earth Frameworks Reviewed

The Earth is not a sphere, but it is pretty close to one. 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. Let's review some terminology and frameworks that will be used during the course of this book:

### Cartography

Figure 1.5.1 shows the Geographic coordinate system for the Earth. Meridians, which are north-south lines on a globe, are given by degrees of longitude. Think of the distance between the north and south poles as long.

- The **prime meridian** passes through Greenwich, Great Britain, at 0 ° longitude.
- East of here (defined as the Eastern Hemisphere, 0 – 180 °E), longitude is positive.
- West of the prime meridian (Western Hemisphere, 0 – 180 °W), longitude is negative.
- The Earth rotates counterclockwise about its axis.

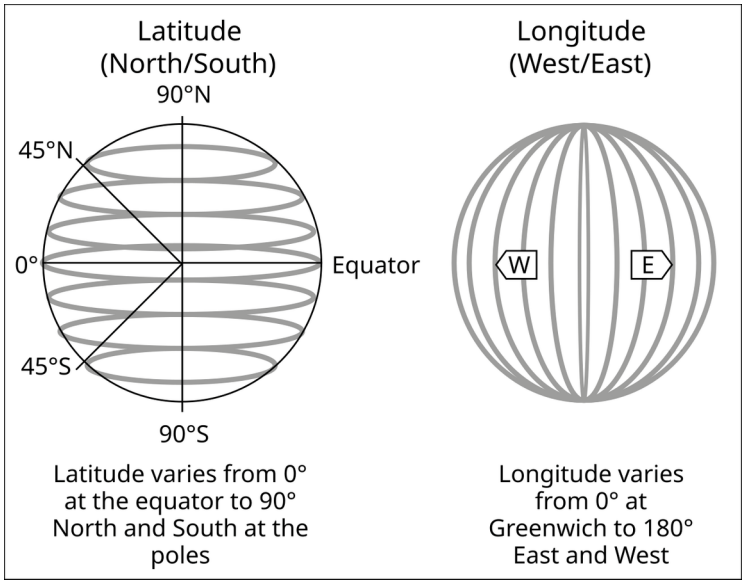



Figure 1.5.1: Longitude lines are perpendicular to, and latitude lines parallel to, the Equator. (Public Domain; Fedstats via Wikimedia Commons)

Parallels, which are east-west lines on a globe, are given by degrees of 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,
- Latitudes north of the Equator are positive (Northern Hemisphere: 0 – 90°N), and
- Latitudes south of the Equator are negative (Southern Hemisphere: 0 - 90°S).

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

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# 1.6: Exercises for Chapter 1

## Chapter 1: Exercises

1. A city like Denver, Colorado (elevation 5,280 ft) will always have a surface pressure lower than Los Angeles, California (elevation of approximately 305 ft in downtown). Why?
2. Carbon Dioxide currently accounts for only 0.042% of the atmosphere (up from 0.038% in Figure 1.1.1). So why do we care so much about it?
3. What is one key feature of each layer of the atmosphere?
4. Of the following statements below, which ones represent Weather and which represent Climate?
   a. Hurricanes are common in Florida, but not in California?
   b. A tornado just moved through downtown Oklahoma City
   c. The average temperature in San Jose, California, is 84°F in July.
   d. Winters in Minneapolis, Minnesota, are very cold.
   e. It was windy last night.

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# 1.7: Accessible Descriptions

## Figure 1.1.1 Composition of the Atmosphere

Pie chart showing the composition of Earth's atmosphere by percentage. Nitrogen accounts for 78.08% and oxygen for 20.95%. The remaining fraction is divided among trace gases, including argon at 0.93%, carbon dioxide at about 0.04%, and very small amounts of neon, helium, methane, krypton, hydrogen, and xenon, each present in fractions of a percent. A more detailed breakdown is provided in Table 1.7.1.

*Table 1.7.1 Percentages of Gases in the Atmosphere*

| Gas | Percentage |
|---|---|
| Nitrogen | 78.08% |
| Oxygen | 20.95% |
| Argon | 0.93% |
| Carbon dioxide | ~0.04% |
| Neon | ~0.0018% |
| Helium | ~0.0005% |
| Methane | ~0.0002% |
| Krypton | ~0.0001% |
| Hydrogen | ~0.00005% |
| Xenon | ~0.000009% |

## Figure 1.1.2 Density vs. Height

Scatter plot showing the relationship between air density (ρ, in kilograms per cubic meter) on the horizontal axis and altitude (z, in kilometers) on the vertical axis. The data points form a smooth curve indicating that density decreases rapidly as altitude increases, with the rate of decrease slowing at higher altitudes. Near ground level, density is highest (just above 1 $kg/m^3$), and it steadily declines toward near zero by around 50 km. The plot includes individual markers connected by a line to show the trend.

## Figure 1.1.3 Pressure vs. Height

Scatter plot showing the relationship between pressure (P, in kilopascals) on the horizontal axis and altitude (z, in kilometers) on the vertical axis. The data points form a smooth curve indicating that pressure decreases rapidly as altitude increases, with the steepest drop occurring near lower altitudes and a more gradual decrease at higher altitudes. Pressure is highest near ground level (around 100 kPa) and approaches very low values by about 50 km. Markers connected by a line illustrate the trend.

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# 2: WARMING THE EARTH AND THE ATMOSPHERE

## Introduction

Outside on a sunny day, you can feel the sun's energy on your skin. You can feel it as *heat*. Standing under the shade of a tree (which blocks the sun's rays) makes a significant difference in temperature and in 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? What does the Earth do with all this energy? And how does it affect our weather?



Figure 2.1: A setting sun with an orange sky shining over a breaking ocean wave (CC BY 2.0; Edmund Garman via Flickr).

### Learning Objectives

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

- Define Air temperature and convert between different scales
- Name and describe the three modes of heat transfer
- Define blackbody radiation and apply Wien's law to determine the maximum emission wavelength
- Describe Earth's surface radiation budget, including shortwave and longwave components
- Describe the cause of seasons on Earth


2.1: Temperature

2.2: Latent Heat

2.3: Heat Transfer

2.4: Electromagnetic waves

2.5: Characterizing Emission

2.6: Radiation Interaction with the Atmosphere

2.7: Surface Energy Balance

2.8: Variations in Radiation

## 2.9: Exercises

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# 2.1: Temperature

## 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 constantly in motion, colliding with one another and with objects. As we increase the temperature of a volume of air, the air molecules speed up, and the frequency of collisions increases. As air temperature decreases, molecules slow down and collide less frequently. What is going on here? Temperature is a relative measure of the average kinetic energy (kinetic meaning motion) of the molecules in a system. Simply put, air temperature is a measure of the average speed of air molecules. Warmer air temperatures indicate that air molecules move faster, whereas cooler temperatures indicate that they move more slowly.

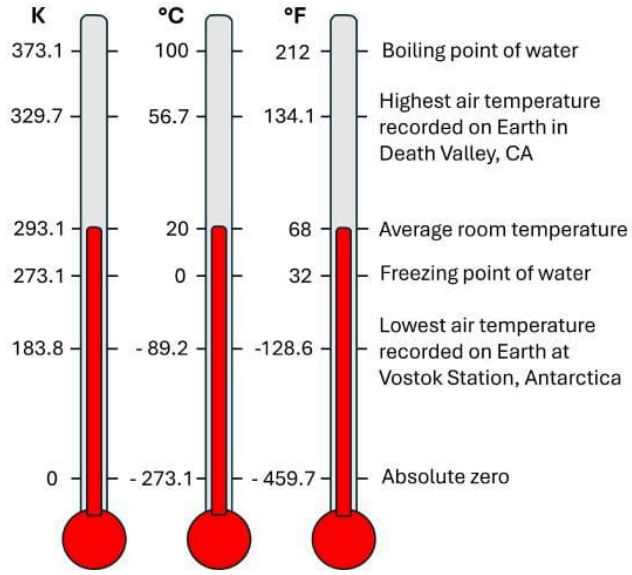



Figure 2.1.1: A comparison between the three major temperature scales: Celsius, Kelvin, and Fahrenheit. (CC BY 4.0; Neel Desai)

*Table 2.1.1: Comparison between the three major temperature scales.*

| Reference point | K (Kelvin) | °C (Celsius) | °F (Fahrenheit) |
|---|---|---|---|
| Boiling point of water | 373.1 | 100 | 212 |
| Highest recorded air temperature (Death Valley, CA) | 329.7 | 56.7 | 134.1 |
| Average room temperature | 293.1 | 20 | 68 |
| Freezing point of water | 273.1 | 0 | 32 |
| Lowest recorded air temperature (Vostok Station) | 183.8 | -89.2 | -128.6 |
| Absolute zero | 0 | -273.1 | -459.7 |

## 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 you probably use in your daily life and what you'll see 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. The highest air temperature ever recorded on Earth is 134.1°F in Death Valley, CA, in July 1913, while the lowest is -128.6°F at the Soviet Vostok Station in Antarctica in July 1983. The average room temperature is around 68°F.

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, as illustrated in Figure 2.1.1. The information from Figure 2.1.1 is also provided in Table 2.1.1. The record for the highest and lowest air temperatures mentioned earlier, °F, correspond to 56.7°C and -89.2°C, respectively. A difference of 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 at rest due to the absence of thermal motion. 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 occurring place in the universe is the Boomerang Nebula, located in the constellation Centaurus and approximately 5,000 light-years from Earth. The temperature is measured at 1 K, only 1°C above absolute zero. A difference of 1 K is equivalent to 1°C, making the conversion linear and straightforward.

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

Based on this, absolute zero is -273.15 °C or -459.7 °F. Keep in mind that 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 around 260-310 K. Within atmospheric sciences, mostly the Kelvin and Celsius scales are used. The Fahrenheit scale is used only when discussing local weather.

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# 2.2: Latent Heat

## Phase Change

Water vapor is an invisible gas that becomes visible when it changes into larger liquid or solid particles. This process of transformation is known as a change of state or phase change. It is the physical, reversible transition of matter among the solid, liquid, and gas states, driven by the absorption or release of heat at constant temperature. Thus, heat must be added or removed from a system during such a phase change, and this heat is called **Latent Heat**. In atmospheric science, we're mostly only concerned with the phase changes of water ($H_2O$).

When heat is added to an ice cube, its temperature increases until it reaches its melting point (0°C). After this point, any additional heat will melt the ice into liquid water, but the temperature of the ice-water system will not change until the phase change is complete. The heat absorbed by the ice-water system from the environment to facilitate the phase change is known as latent heat. You have experienced this as your ice melts in your water glass. It stays cold until all the ice is gone, 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 to transition from liquid to vapor. Energy is absorbed from the environment (your skin) to evaporate the remaining liquid, which cools your skin. 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, and deposition/resublimation) are warming processes because energy is released during the phase change. This is an important distinction, so ensure it is clear before proceeding.

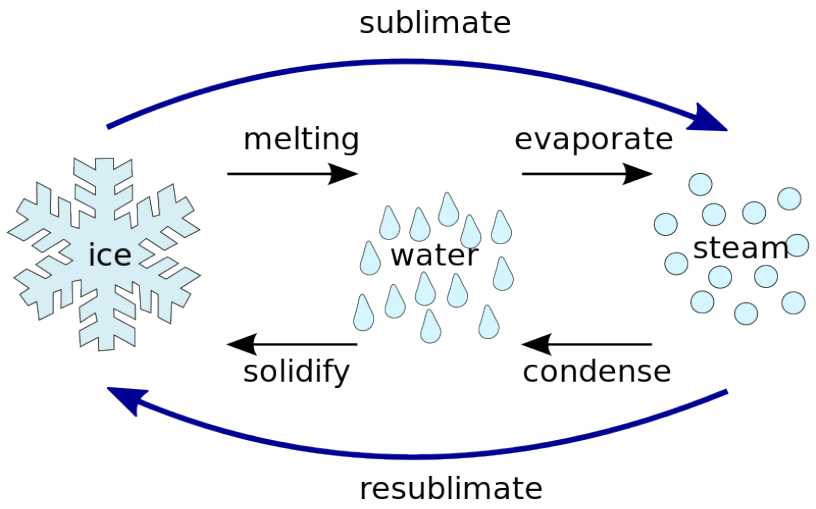



Figure 2.2.1: Phase changes of water. Melting, evaporation, and sublimation are considered cooling processes because they take heat from the environment. Condensation, freezing, and deposition are considered warming processes because, during the phase change, energy is released into the environment. (CC BY-SA 3.0; Synkizz via Wikimedia Commons).

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

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# 2.3: Heat Transfer

## Modes of Heat Transfer

A critical example of kinetic energy is **thermal energy**, which arises from the microscopic movement of many molecules within a system. We discussed how *temperature* is a measure of the average speed of atoms and molecules in a system. The faster the molecules move, the higher the temperature. When two objects (solid, liquid or gaseous) with different temperatures either come into contact or are placed close to each other, this thermal energy can be transferred from the object with higher energy to the object with lower energy. The transfer of thermal energy due to a temperature difference between two objects is known as **heat**. Heat is a form of energy in transit, and once transferred, it is stored as internal energy. Heat, being a form of energy, is given in standard units of Joules (J).

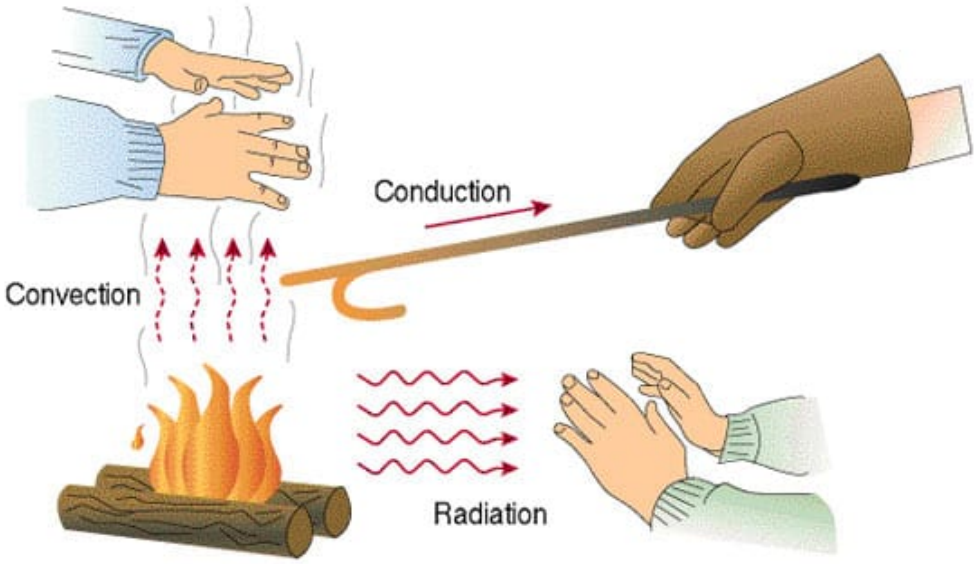



Figure 2.3.1: Conduction, convection, and radiation are all methods of heat transfer. When you're sitting by a fireplace, you feel the heat via radiation. Warm air and smoke rise from the fire via convection, and a marshmallow stick or a hot object heats up by conduction. (CC BY-SA 4.0; Kmecfiunit via Wikimedia Commons).

Figure 2.3.1 illustrates the three primary methods of heat transfer: *conduction*, *convection, and radiation*. However, there is a fourth mode of heat transfer, which is very important for the atmosphere: the absorption or release of *latent heat*.

## Conduction

If you hold a metal rod over a fire, you feel the energy from the fire is immediately transferred to your hand, causing you to feel a burn. This is an example of **conduction**: energy is directly transferred through a substance without the movement of material. The molecules of the rod directly above the fire absorb heat from the flame and begin to vibrate faster due to the additional kinetic energy. This energy is transferred to the neighboring lower-energy molecules through vibrations until it reaches your fingers. Contact with these vibrating rod molecules causes the molecules in your fingers to begin vibrating, in a sensation that we perceive as heat. Certain materials are better conductors of heat than others. Metals, for example, conduct heat very efficiently, while air, which acts as an insulator, is a very poor conductor of heat. This is why air and other poor conductors, such as styrofoam, are often used as insulators. However, that does not mean that air cannot transfer thermal energy.

## Convection

Another type of energy transfer is convection, which occurs in liquids and gases (both fluids) because their molecules can move freely, allowing currents to form naturally. **Convection** refers to the heat transfer by the macroscopic movement of a fluid. This occurs because lower-density fluid tends to rise over higher-density fluid, which sinks under the force of gravity, thereby

transferring heat within the fluid. In Figure 2.3.2, 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 denser and sinks toward the bottom under gravity.

Convection occurs commonly in the atmosphere. The Sun heats the Earth's surface unevenly, resulting in certain regions being warmer than others. The air above the warmer regions receives energy from the ground via conduction. Since air is a poor conductor, it cannot transfer this energy to its neighboring molecules via conduction. Instead, the warm air rises due to positive buoyancy and is replaced at the surface by sinking cooler air, similar to the beaker in Figure 2.3.2.

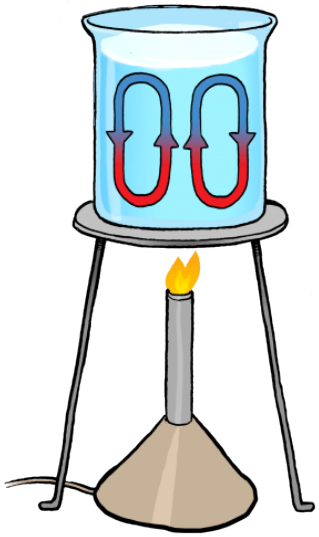


Figure 2.3.2: Convection within a fluid that is heated from the bottom. Warm fluid rises up near the middle, and is replaced at the bottom by colder fluid from the sides of the beaker. (CC BY 2.0; Siyavula Education via Flickr).

## Radiation

When you stand outside in the Sun, you feel warm. Similarly, if you're sitting next to a fireplace or a bonfire, you feel its warmth without having to place your hands in it. This transfer of energy from the Sun or the fireplace is called *radiant energy* or radiation. **Radiation** is the transfer of energy via electromagnetic waves. Radiant energy can be transferred to an object without the space in between being heated. 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 $km \cdot s^{-1}$. In the atmosphere, this is the most important form of energy transfer.

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# 2.4: Electromagnetic waves

## Waves

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, like the one in Figure 2.4.1. Wave periods of 12, 14, or longer seconds are ideal. Depending on whether you are a beginner or an advanced rider, you might prefer a wave face height of 2 to 30 feet. The wave period is 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 more complex in surfing, as it is measured at wave break. In the case of electromagnetic energy, the waves do not break.



Figure 2.4.1: Waves at a beach are an example of a commonly experienced wave phenomenon. (Public Domain; NOAA via Wikimedia Commons).

## Electromagnetic Radiation

Radiation can be considered in two ways: as electromagnetic waves or as photons. In atmospheric science, we generally consider radiation as a wave rather than a photon. **Electromagnetic radiation** is a form of energy produced by oscillating electric and magnetic fields and is known by various names depending on its wavelength. For example, you've probably heard of radio waves and X-rays, where the primary difference between the two forms of electromagnetic radiation is the wavelength of the waves in each. Figure 2.4.2 shows the relationship between wavelength, frequency, and temperature. Colder objects radiate at lower frequencies and longer wavelengths, and warmer objects radiate at higher frequencies and shorter wavelengths. This relationship holds across all scales, from atomic nuclei to planets, and across temperatures from near absolute zero to millions of Kelvins.

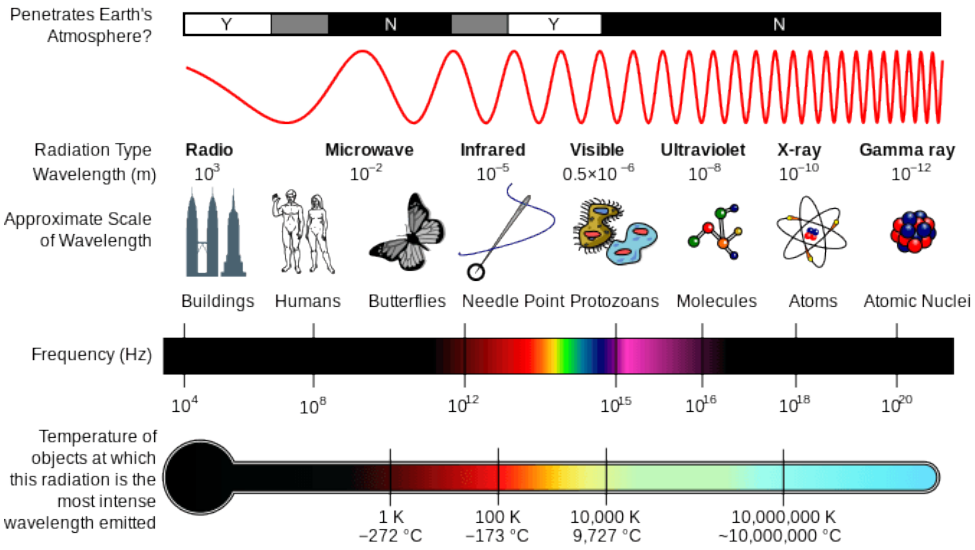



Figure 2.4.2: The electromagnetic energy spectrum, including whether it penetrates the Earth's atmosphere, wavelength visual (red line), radiation type, wavelength value in m, wavelength scale comparison, frequency in Hz, and temperature of black body objects that emit at that wavelength of radiation. Table 2.4.1 provides the information in a tabular form. (CC BY-SA 3.0; Inductiveload via Wikimedia Commons).

*Table 2.4.1: The relationship between types of electromagnetic waves, including their wavelength, frequency, temperature, ability to penetrate Earth's atmosphere, approximate size of wavelength, and equivalent temperature.*

| Region of the Spectrum | Approximate Wavelength (meters) | Frequency (Hz) | Penetrates the Earth's atmosphere? | Approximate size of wavelength | Temperature Equivalent (Kelvin) |
|---|---|---|---|---|---|
| Radio | > 1 meter | < 300 MHz | Yes | Radio towers, buildings | ~ 1 |
| Microwave | 1 m to 1 mm | 300 MHz – 300 GHz | No | Butterflies | ~10 |
| Infrared (IR) | 1 mm to 700 nm | 300 GHz – 400 THz | Partly | Needle point | ~100 |
| Visible Light | 700 nm to 400 nm | 400 THz – 750 THz | Yes | Protozoans | $\sim 10^3$ |
| Ultraviolet (UV) | 400 nm to 10 nm | 750 THz – 30 PHz | Mostly no | Molecules | $\sim 10^4$ |
| X-rays | 10 nm to 0.01 nm | 30 PHz – 30 EHz | No | Atoms | $\sim 10^7$ |
| Gamma rays | < 0.01 nm | > 30 EHz | No | Atomic nuclei | $\sim 10^8$ |

Radio waves have the lowest energies, the longest wavelengths (100 km to 1 m), and the lowest frequencies of any type of electromagnetic radiation. Thus, they occupy the left-most position in Figure 2.4.2. Radio waves are essential for wireless technology, including AM/FM radio broadcasting, cellular telephone communications, television, radar, and satellite communication. As we move toward higher-energy waves, Microwaves are next, with slightly higher energy and shorter wavelengths (1 mm to 1 m) than radio waves. Certain microwave frequencies (such as those used in microwave ovens) are absorbed by water molecules, causing them to vibrate and generate heat. But the energy in these waves is still too low to cause bodily harm. Next is Infrared radiation, with wavelengths from 1 mm to 700 nm, emitted by ordinary objects such as rocks, the Earth, and even humans. These waves are often used in fibre optic communication and remote controls. Next up is visible light, with wavelengths ranging from 700 to 400 nm. It comprises only a tiny fraction of the EM spectrum, yet it accounts for everything we see and nearly 47% of the energy Earth receives from the Sun. Ultraviolet radiation (UV) is located next in the electromagnetic spectrum, with wavelengths shorter than those of visible light (400-100 nm). We have finally reached a region of the spectrum with sufficient energy to damage objects and human skin. The Sun emits UV radiation, most of which is blocked by the Ozone layer and is prevented from reaching the Earth's surface. Still, some UV radiation reaches the surface, which is why wearing sunscreen is important if you're going to be out in the Sun. Next up are X-rays with significantly higher energies and shorter wavelengths

(10 to 0.1 nm). They possess significant penetrating power, enabling them to pass through soft tissues while being absorbed by denser tissues such as bone, making them essential for medical and dental imaging. However, prolonged or repeated exposure to X-rays can cause significant tissue and DNA damage, leading to cancer and other health risks. Gamma rays occupy the far right side of the EM spectrum and have the highest energies and the shortest wavelengths of any other EM waves. They are generated by atomic nuclei during radioactive decay, nuclear fission/fusion, or violent astronomical phenomena such as supernovae and black holes. These high-frequency, high-energy radiations are among the most dangerous to humans, and even mild exposure can cause significant cellular damage.

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# 2.5: Characterizing Emission

## Wavelength, Frequency, and Speed

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 " $c$ ", approximately $3\times10^8$ $m{\cdot}s^{-1}$). The waves have a wavelength ($\lambda$) given by the distance from one wave crest to the next. The waves also have a frequency ($v$), which is the number of repeated wave occurrences in a specified period of time. The unit for frequency is Hertz (cycles per second).

$$Wavelength \times Frequency = Speed \tag{2.5.1}$$

$$\lambda \times v = c$$

With variables for wavelength and frequency, we can define any type of wave with great clarity. But 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. To estimate the amount of radiant energy an object emits, a common simplification is needed: we assume the object behaves as a blackbody. A **blackbody** is an object that emits and absorbs the maximum amount of radiation possible given its temperature. The object doesn't have to be black; it just has to absorb all wavelengths and emit all the radiation back, without storing any energy. Both the Earth and the Sun are considered blackbodies for most practical purposes.

## 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. Figure 2.5.1 shows multiple Planck function curves for various black-body temperatures. The red, green, and blue lines represent objects at 3000 K, 4000 K, and 5000 K, respectively. As a black body becomes hotter, it emits at shorter wavelengths and at higher intensities. Higher-temperature curves are taller and broader, peaking at shorter wavelengths.

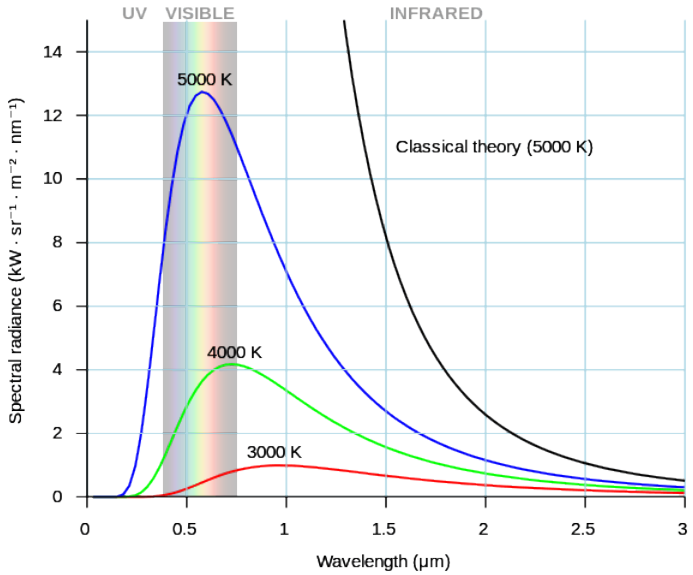



Figure 2.5.1: The spectral radiance (y-axis, emitted radiation) versus the wavelength (x-axis) of radiation emitted from black bodies of various temperatures. (Public Domain; Darth Kule via Wikimedia Commons).

Table 2.5.1 provides the relevant information from Figure 2.5.1

*Table 2.5.1. Approximate peak wavelength, region, and intensity for objects of various temperatures.*

| Temperature (K) | Peak wavelength (μm) | Peak region | Relative peak intensity |
|---|---|---|---|
| 5000 | ~0.5 | Visible | Highest |
| 4000 | ~0.7 | Visible | Moderate |
| 3000 | ~1.0 | Infrared | Lowest |

## Wien's Law

Planck's function tells us the amount of emitted radiation and the wavelengths over which it is emitted, given the temperature of the black body. We can use another law to determine the maximum wavelength that a black body can emit. **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 2898.

$$\text{Maximum Wavelength} = \frac{a}{\text{Temperature}} \tag{2.5.2}$$

$$\lambda_{\max} = \frac{2898(\mu m \cdot K)}{T(K)} \tag{2.5.3}$$

Using Wein's equation to find the wavelength of electromagnetic radiation yields an answer in microns (μm). One micron is equal to $10^{-6}$ meters.

### Example 2.5.1: 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:

$$\lambda_{\max} = \frac{2898}{T}$$

$$\lambda_{\max} = \frac{2898}{5800K} = 0.499$$

This radiation, with a wavelength of 0.499 μm, falls within the visible-light range, which is typically defined as 0.5 × 10^-6 m.

## Stefan-Boltzmann Law

One last fundamental 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 shows that the hotter an object is, the more energy it radiates per unit area.

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

$$E = \sigma * T^4 \tag{2.5.4}$$

Equation 2.6.2 shows that the radiance emitted per unit area is proportional to the blackbody temperature raised to the fourth power. This relationship is extremely important. It shows that the amount of radiation emitted depends heavily on temperature, so small temperature fluctuations can lead to large changes in emittance. The Stefan-Boltzmann constant is $5.67\times10^{-8}$ $W{\cdot}m^{-2}{\cdot}K^{-4}$ and the symbol sigma, $\sigma$, is used.

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

## Solar versus terrestrial radiation

The Sun's average temperature is above 5,000 K, while the Earth's average temperature is in the range 210-310 K. This means that the Sun and Earth radiate energy very differently. This follows directly from the electromagnetic energy spectrum and the respective temperatures of the Sun and Earth, as illustrated in Figure 2.5.2. The Sun emits higher intensity radiation at shorter wavelengths (0.2 - 4 μm), with a peak in the visible spectrum around 0.5 μm. This is because it has a higher temperature than Earth, and Planck's curve for higher temperatures peaks at shorter wavelengths. The Earth emits lower-intensity radiation at longer wavelengths (3-100 μm), with a peak in the infrared spectrum around 10 μm. Thus, in colloquial terms, the Sun emits *shortwave radiation,* and the Earth emits *longwave radiation*.

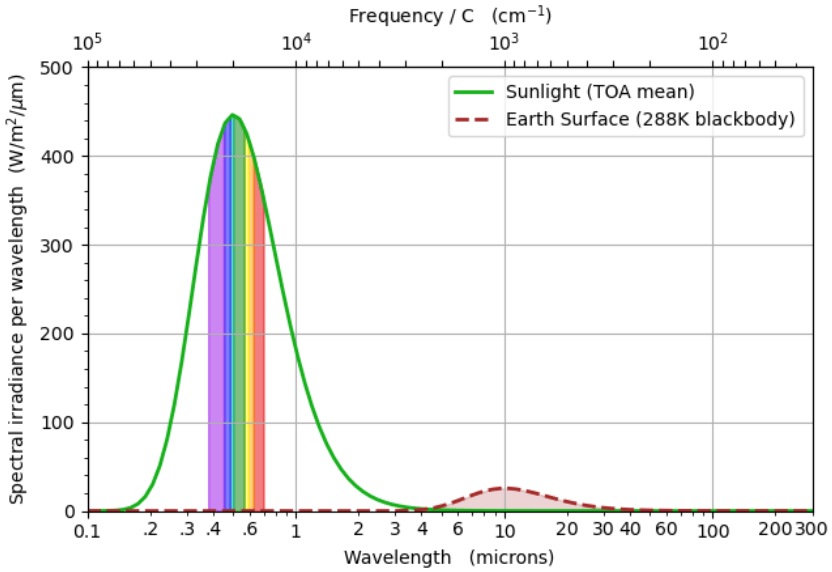



Figure 2.5.2: The spectral irradiance or intensity of Solar and terrestrial radiation is shown with respect to wavelength. The Sun emits higher-intensity radiation at shorter wavelengths, while the Earth emits lower-intensity radiation at longer wavelengths. (CC BY-NC-SA 4.0; Robert Wentworth via Climate Puzzles)

The relevant information from Figure 2.5.2 is provide in Table 2.5.2.

*Table 2.5.2. Peak characteristics for the Sun vs. Earth*

| Source | Peak wavelength (μm) | Peak irradiance (W/m²/μm) | Region |
|---|---|---|---|
| Sunlight (TOA mean) | ~0.5 | ~450 | Visible |
| Earth surface (288 K) | ~10 | ~30 | Infrared |

So, what happens to all this high-energy solar radiation that impacts the Earth? 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 Earth absorbs all incoming radiation from the Sun and emits as much as possible given its temperature. In practice, it is a little bit more complicated than this.

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# 2.6: Radiation Interaction with the Atmosphere

## Albedo

Not all incoming solar radiation is absorbed by Earth because it is not a *perfect* blackbody. 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.

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

An object that reflects all incident radiation has an albedo of 1. On the other hand, an object that absorbs all the radiation and doesn't reflect anything will have an albedo of 0.

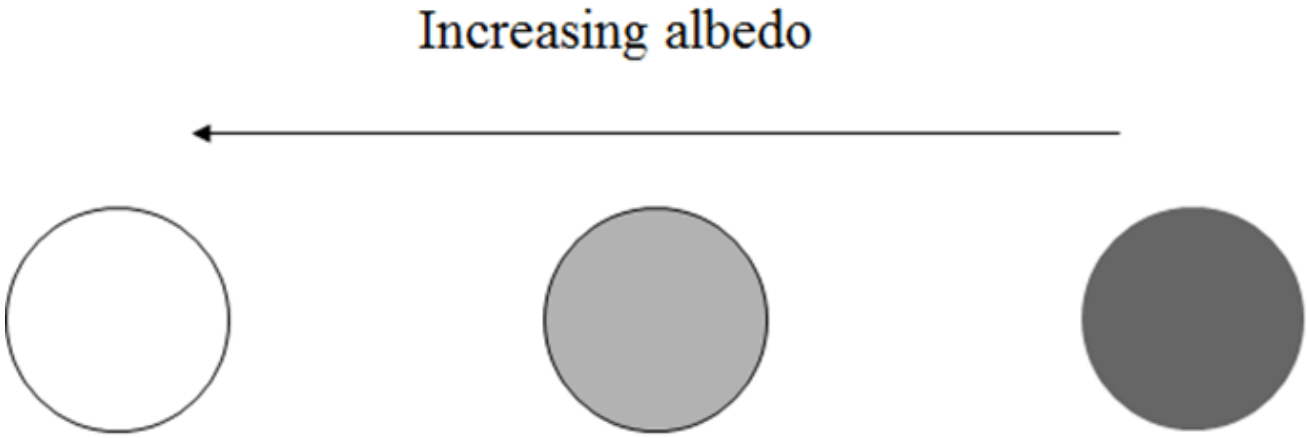



Figure 2.6.1: Relationship between Albedo and the darkness of objects. Lighter colors have higher albedo, whereas darker colors have lower albedo. (CC BY 4.0; Alicia Mullens)

The simplest way to think of albedo is in terms of an object's color. White objects are highly reflective and have a high albedo because the amount of reflected light is much greater than the amount of incident light, as illustrated in Figure 2.6.1. You'll know this if you've been on a white-sand beach or a snowfield and felt the brightness of the environment. Following the same logic, dark surfaces absorb light; therefore, black, brown, or dark green colors typically have 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 lines demarcating parking spaces.

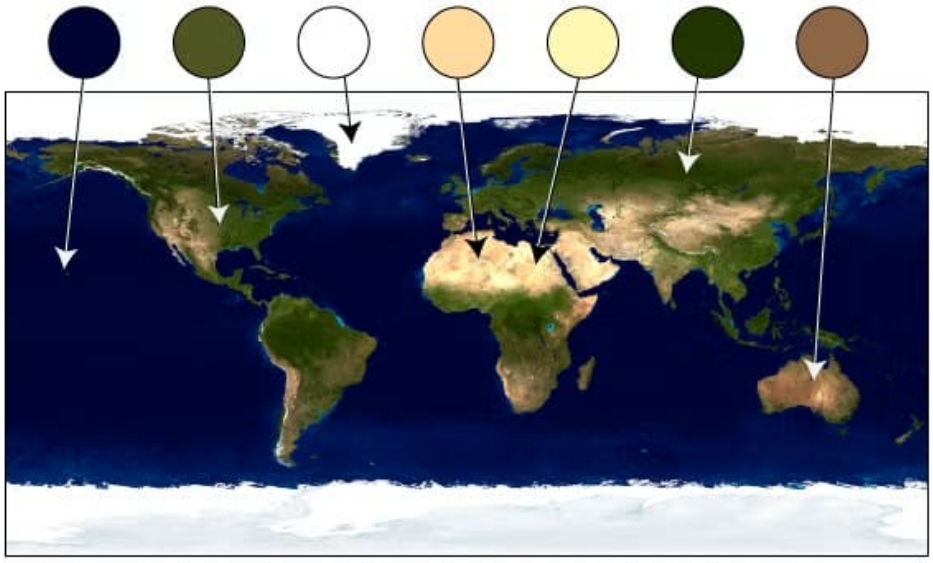


Figure 2.6.2: A sampling of Earth's colors. Snow-covered regions and deserts appear more reflective than forests or oceans. (CC-BY-NC 4.0; UCAR SciEd via NASA)

The Earth offers a wide range of colors for incoming solar radiation as illustrated in Figure 2.6.2. Fresh snow has a high albedo of nearly 0.9. Thus, Earth's snow-covered high-latitude regions have albedo values ranging from 0.75 to 0.95. Clouds also reflect sunlight and can have albedos ranging from 0.3 to 0.9, depending on their brightness. Meanwhile, forests and grasslands are darker with albedo values ranging from 0.1 to 0.3. Open oceans are even darker, with extremely low albedo values of 0.06-0.1. Note that radiation reflected from an object does not warm the object. Earth has an average planetary albedo of approximately 0.3 (30%).

## Selective absorbers

While some components of the Earth-atmosphere system, such as snow and clouds, reflect radiation, others absorb only specific wavelengths. These are called *selective absorbers;* examples include certain gases in Earth's atmosphere. Figure 2.6.3 shows the spectral intensity of incoming solar radiation (UV and visible in red) and outgoing thermal radiation (IR in blue). The subsequent panels show the total percentage of radiation absorbed and scattered by the atmosphere as a function of wavelength, decomposed by primary greenhouse gases. **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.

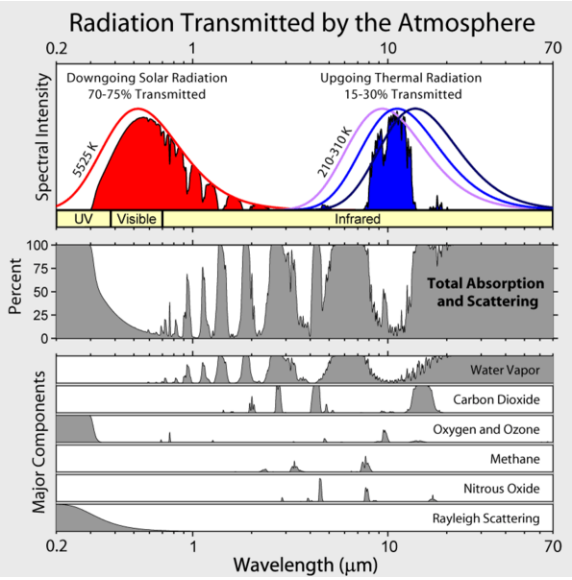



Figure 2.6.3: Radiation transmitted by the atmosphere as a function of wavelength. Downgoing solar radiation is on the left, and upgoing thermal radiation is on the right, with the percentage of radiation absorbed and scattered by the Earth's atmosphere. Finally, the percentages of radiation absorbed and scattered by the major greenhouse gases are shown in the subsequent panels. Table 2.6.1 provides details in a tabular form. (CC BY-SA 3.0; Global Warming Art via Wikimedia Commons).

*Table 2.6.1: Radiation transmitted by the atmosphere as a function of wavelength, source, and percentage absorbed.*

| Type of Radiation | Wavelength Range (μm) | Radiation Type | Absorption and Scattering (%) | Notes |
|---|---|---|---|---|
| Downgoing solar radiation | 0.2 – 0.3 | Ultraviolet (UV) | 100 | Strong absorption by ozone |
| | 0.3 – 1 | Visible light | Less than 25% | High transparency allows solar incoming radiation to reach Earth |
| | 1 – 3 | Near-infrared | ~ 50% | Some absorption by water vapor |
| Upgoing terrestrial radiation | 3 – 8 | Mid-infrared | ~ 75% | Strong absorption by water vapor & $CO_2$ |
| | 8 – 14 | Infrared "window" | Less than 25% | High transparency allows outgoing radiation to escape to space |
| | > 15 | Far-infrared | 100% | Strong absorption by water vapor & $CO_2$ |

In Figure 2.6.3 and Table 2.6.1, observe the total percentage of radiation that is absorbed and scattered by the Earth's atmosphere. You'll notice two large gaps. These are called **Atmospheric Windows,** where radiation can pass through the atmosphere without interacting with greenhouse gases. Particularly notable is the *Optical window*, between 0.3 and 1 μm, which allows most of the visible portion of the electromagnetic energy spectrum, as well as some infrared solar energy, to pass through the atmosphere unimpeded. Meanwhile, various atmospheric gases absorb a large portion of the infrared spectrum. This suggests that Earth's atmosphere is nearly transparent to shortwave radiation but strongly absorbs infrared radiation. This will be important for the climate factors discussed in a later chapter. Notice that there is another Atmospheric Window in the infrared spectrum, between 8 and 14 μm. This is known as the *infrared atmospheric window,* which allows some of Earth's upgoing longwave radiation to pass through the atmosphere into space, thereby cooling the Earth. While clouds are not included in Figure 2.6.3, they can absorb some of this upgoing radiation between 8 and 11 μm.

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# 2.7: Surface Energy Balance

## Radiative Equilibrium Temperature

The average temperature of a planet depends on the balance between incoming radiation and outgoing radiation. If incoming radiation exceeds outgoing radiation, a planet will continuously warm. If outgoing radiation exceeds incoming radiation, a planet will constantly cool. Eventually, a planet tends toward *radiative equilibrium*, in which the energy of outgoing radiation equals that of incoming radiation. The average temperature at which this occurs is called the **radiative equilibrium temperature**.

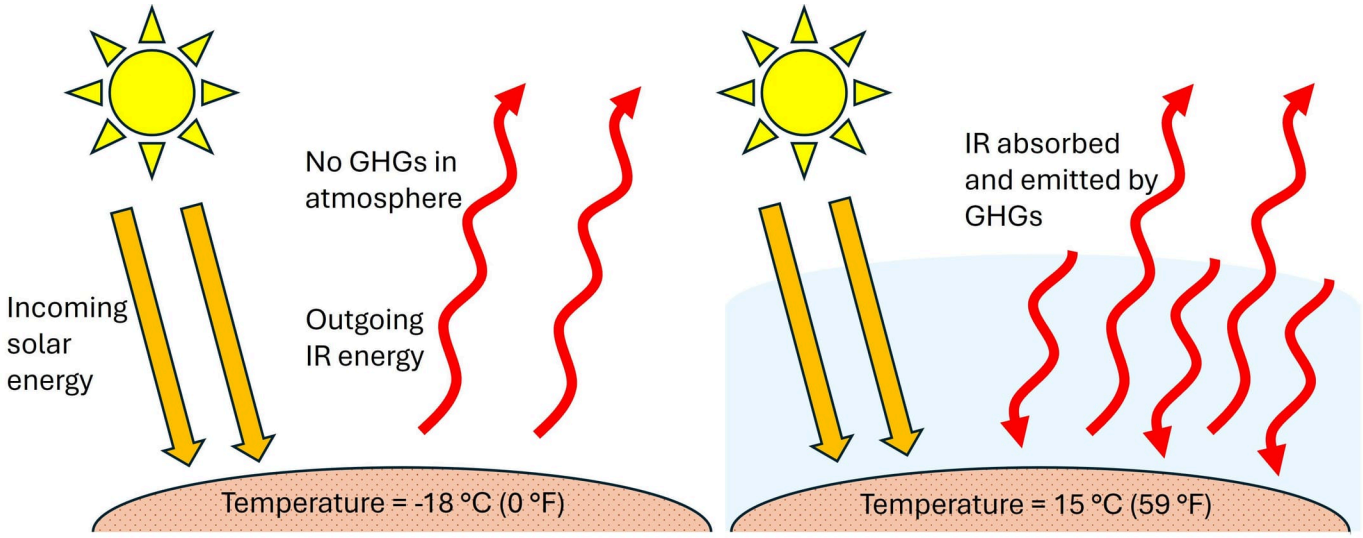



Figure 2.7.1: Incoming solar energy warms the Earth's surface via shortwave radiation, and the Earth emits all the energy back to space via Infrared (IR) or longwave radiation. Without greenhouse gases (GHGs) in the atmosphere, the Earth's radiative equilibrium temperature would have been -18 °C. However, GHGs in the atmosphere absorb outgoing IR radiation and radiate some of it back toward the surface, increasing the surface temperature to a more comfortable 15 °C. (CC BY 4.0; Neel Desai)

If the Earth didn't have an atmosphere, it would be just a rock in space at approximately 150 million kilometers (93 million miles) from the Sun, as illustrated in Figure 2.7.1. At this distance, half of the Earth will always receive Solar radiation and the other half will be in darkness. Since Earth's temperature is above 0 K, it continuously emits radiation out to space. During the day, the incoming solar radiation is much larger than the outgoing terrestrial radiation, resulting in warming the Earth. However, at night, there is no incoming solar radiation, and outgoing radiation cools the Earth. The combined effect of incoming and outgoing radiation would result in the Earth's radiative equilibrium temperature being about 255 K (-18°C, 0°F). This temperature is much lower than the Earth's observed average surface temperature. This is because the Earth isn't just a rock floating in space. It has an atmosphere and a surface with varying degrees of reflectivity. When considering the Earth's atmosphere, including albedo and the greenhouse gases that are transparent to incoming shortwave radiation but absorb most outgoing longwave radiation, the Earth's Radiative equilibrium temperature increases to 288 K (15°C, 59°F).

## Earth's Annual Energy Balance

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. The interaction of atmospheric gases in Earth's atmosphere is an important part of its energy balance. Figure 2.7.2 illustrates that clouds reflect some of the incoming solar radiation, the Earth's surface reflects some, but most is absorbed by the Earth's surface or the atmosphere.

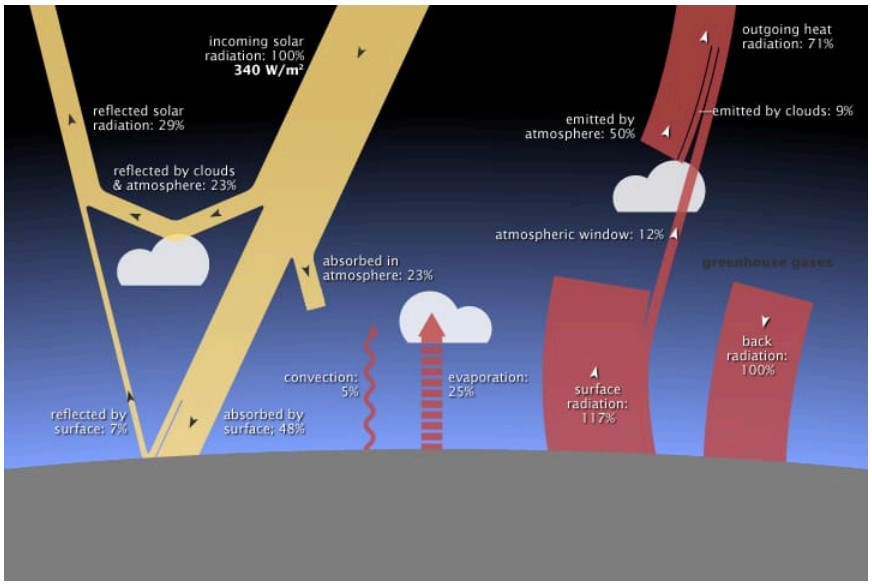



Figure 2.7.2: On average, 340 watts per square meter of solar energy arrives at the top of the atmosphere. Earth returns an equal amount of energy to space by reflecting some incoming light and by radiating heat. Most solar energy is absorbed at the surface, while most heat is radiated back to space by the atmosphere. (Public Domain; Climate and Earth's Energy Budget via NASA).

The details and percentages from Figure 2.7.2 are provided in Table 2.7.1

*Table 2.7.1: A breakdown of Earth's energy budget.*

| Process / Component | Percentage (%) | Notes |
|---|---|---|
| Incoming solar radiation | 100 | 340 W/m² total input |
| Reflected solar radiation (total) | 29 | Returned to space |
| – Reflected by clouds & atmosphere | 23 | Major reflection component |
| – Reflected by surface | 7 | Surface albedo |
| Absorbed solar radiation (total) | 71 | Energy retained in Earth system |
| – Absorbed by atmosphere | 23 | Includes gases and clouds |
| – Absorbed by surface | 48 | Land and oceans |
| Surface radiation (emitted upward) | 117 | Infrared emission from surface |
| Back radiation (toward surface) | 100 | Re-emitted by greenhouse gases |
| Emitted by atmosphere | 50 | Outgoing to space |
| Emitted by clouds | 9 | Outgoing to space |
| Atmospheric window (direct escape) | 12 | Surface radiation escaping directly |
| Outgoing heat radiation (total) | 71 | Balances absorbed solar energy |
| Evaporation (latent heat transfer) | 25 | Surface to atmosphere |
| Convection (sensible heat transfer) | 5 | Surface to atmosphere |

Aerosols, water vapor, and ozone in our atmosphere directly absorb 23% of incoming solar energy. Nearly 7% of the sunlight is reflected from the Earth's surface, and clouds reflect an additional 23%. The surface absorbs the rest. To maintain a constant radiative equilibrium temperature, the Earth must radiate away all the energy it receives. Evaporation and convection account for 25% and 5% of the incoming solar energy transferred from the surface to the atmosphere, respectively. However, if you calculate the energy emitted by the Earth using the Stefan-Boltzmann Law**,** with an average surface temperature of 288 K, you will find that the Earth's surface emits nearly 117% of the incoming solar energy, of which only 48% is actually absorbed by the surface. Where did this additional energy come from?

# The Greenhouse Effect

The **greenhouse effect** is the result of greenhouse gases in the atmosphere absorbing outgoing infrared radiation from Earth's surface and re-emitting it. 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, it has, in fact, made Earth habitable. In Figure $2.7.2$, you'll notice that the greenhouse gases in our atmosphere radiate a full 100% of solar energy in the form of back radiation to the surface. This additional energy raises Earth's radiative equilibrium temperature to 288 K, and the Earth's surface must re-emit more than 48% of the absorbed solar radiation back into space.

Over the past 100 to 150 years, atmospheric ($CO_2$) concentration has increased rapidly from 313 parts per million (ppm) in 1960 to 424 ppm in 2024. Human activities such as burning fossil fuels for energy and transportation, cement production, and tropical deforestation are the primary drivers of this rapid increase in atmospheric ($CO_2$) concentrations. This has led to increased absorption of outgoing terrestrial longwave radiation by the atmosphere, resulting in an average warming of 2.62 °F (1.35 °C) since the pre-industrial era (1850-1900). This strengthening of the greenhouse effect due to additional greenhouse gases from human activities is known as the *enhanced greenhouse effect*. According to the 2017 U.S. Climate Science Special Report, if yearly emissions continue to increase rapidly, as they have since 2000, models project that by the end of this century, global temperature will be at least 5 °F warmer than the 1901-1960 average, and possibly as much as 10.2 °F warmer.

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# 2.8: Variations in Radiation

## Radiation Variations Across Earth

The Earth maintains a radiative balance between the energy it receives from the Sun, the energy in its atmosphere, and the energy it radiates into space. But that does not mean that different regions of the Earth receive the same amount of energy every day. The energy received by various parts of the Earth depends on the season and latitude.

## Seasonal Changes

The Earth revolves around the sun in an elliptical orbit in 365.25 days, or one year. Since the orbit is elliptical, the actual distance between the Earth and the sun varies throughout the year. It is a common misconception that this is the reason for our seasons. However, that is not true. The Earth is closest to the sun in January (147 million km), while it is furthest from the sun in July (152 million km). This would suggest that summer occurs in January and winter in July, but that's not how we experience our seasons in the Northern Hemisphere.

**The Earth's tilt is responsible for our seasons.** If the Earth did not have a tilt, we would not have seasonal changes. The Earth is tilted by 23.5° on its axis**.** Earth's tilt with respect to space does not change throughout the year; thus, the Northern Hemisphere is tilted toward the sun in June and away from it in December, as illustrated in Figure 2.8.1. The tilt does not significantly change the distance of each hemisphere from the sun. But it does change the angle at which sunlight strikes the ground. This is called the *angle of insolation*.

Why should this *angle of insolation* affect our seasons? To understand this, take a piece of paper and a flashlight. Shine the flashlight directly onto the paper so you see a circle illuminated. All the light from the flashlight is in that circle. Now, slowly tilt the paper so that the circle elongates into an ellipse. All the light is still in that ellipse, but the ellipse is spread out over more paper. The *density* of light drops. In other words, the amount of light per square centimeter drops (the number of square centimeters increases, while the total amount of light stays the same). The same is true on the earth. When the sun is overhead, the light is falling straight on you, and so more light (and more heat) hits each square centimeter of the ground. When the sun is lower in the sky, the light gets more spread out over the surface of the earth, and less heat (per square centimeter) can be absorbed. Since the Earth's axis is tilted, the sun is higher when you are on the part of the Earth where the axis points more towards the sun, and lower on the part of the Earth where the axis points away from the sun. The angle of the Sun's rays and the length of the day change because of the Earth's tilt.

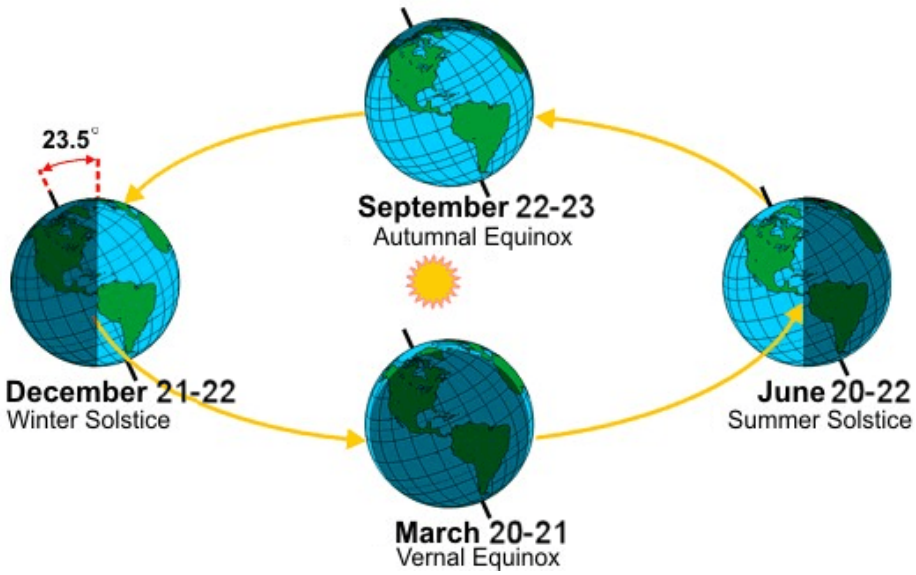



Figure 2.8.1: Due to the Earth's tilt, the northern hemisphere is tilted towards the Sun in June, and tilted away from it in December. The northern hemisphere experiences summer when it is tilted towards the Sun and winter when it is tilted away. (Public Domain; The Seasons via NWS).

During the summer, each hemisphere faces the sun. For example, during Northern Hemisphere summer in June, the sun shines more directly on the Northern Hemisphere than on the Southern Hemisphere. On June 20-22, the sun is directly above the Tropic of Cancer, at 23.5° north latitude. This day, known as the **Summer Solstice**, marks the longest day of the year in the Northern Hemisphere and the shortest in the Southern Hemisphere. While it seems like a small change, the angle of insolation and the amount of solar radiation absorbed (per area) vary significantly throughout the year. This is because, in addition to the Sun's angle, the Earth's position changes the length of the day throughout the year. On September 22-23 and on March 20-21, the sun is directly above the equator. These days are called equinoxes (equal days), since both hemispheres enjoy nearly 12 hours of daylight and 12 hours of night. March 20-21 is called the **Vernal (Spring) Equinox**, while September 22-23 is called the **Autumnal (Fall) Equinox**. On December 21-22, the sun is directly above the Tropic of Capricorn, at 23.5° south latitude. This day, known as the **Winter Solstice**, is the shortest day of the year in the Northern Hemisphere but the longest in the Southern Hemisphere. Figure 2.8.2 depicts 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.

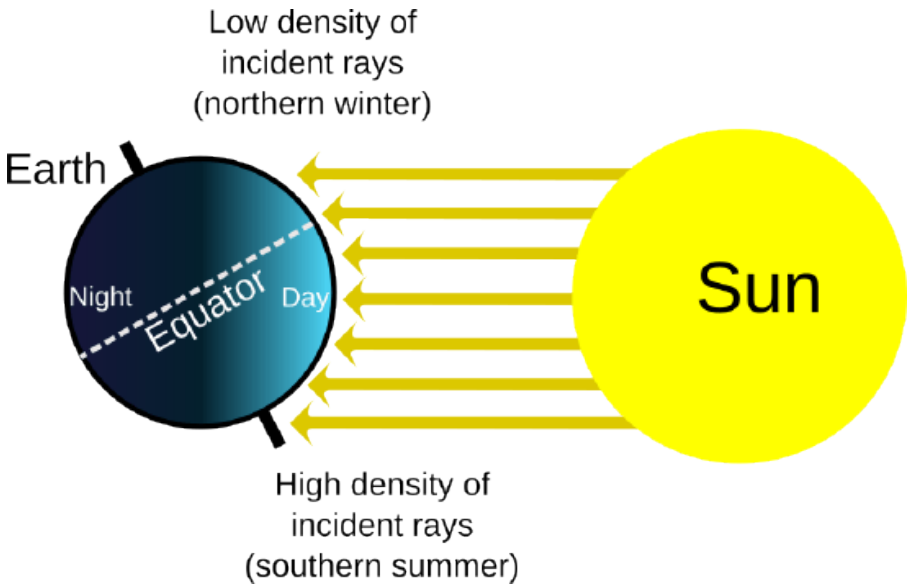



Figure 2.8.2: Distribution of solar rays on Earth during the Northern Hemisphere winter. The Southern Hemisphere, experiencing summer, receives high-intensity incident solar radiation, while the Northern Hemisphere, experiencing winter, receives low-intensity solar radiation. (Public domain; Rhcastilhos via Wikimedia Commons).

# Radiation vs Latitude

In the “summer hemisphere,” the combination of more direct sunlight and longer days means the pole can receive more incoming sunlight than the tropics, but in the winter hemisphere, it gets none. Even though illumination increases at the poles in summer, bright white snow and sea ice reflect a significant portion of incoming light, reducing the potential for solar heating. Differences in albedo and solar illumination at different latitudes lead to net heating imbalances across the Earth system. At any place on Earth, the net heating is the difference between the amount of incoming sunlight and the amount of heat radiated by the Earth back to space. Figure 2.8.3 depicts the relationship between the average annual incoming solar and outgoing terrestrial radiation with latitude. The blue line indicates the net shortwave radiation received by the Earth. In contrast, the red line indicates the net longwave radiation emitted annually by the Earth and the atmosphere. In the tropics, there is a net energy surplus because the amount of sunlight absorbed exceeds the amount of heat radiated. In the polar regions, however, there is an annual energy deficit because the amount of heat radiated to space exceeds the amount absorbed from sunlight.

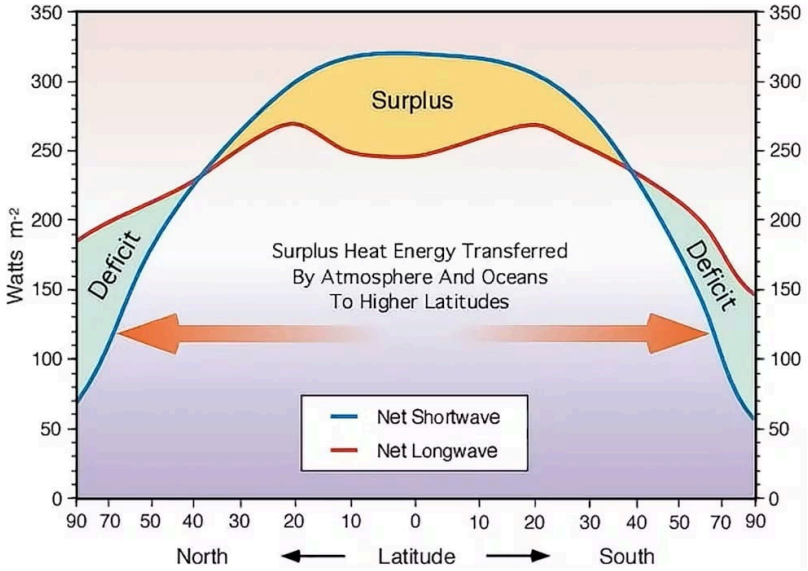



Figure 2.8.3: Latitudinal variation in Earth's radiation balance. The tropics receive surplus heat energy, while the polar regions have an energy deficit. The surplus heat energy is transferred from the tropics to higher latitudes by the atmosphere and the oceans. (CC BY-NC 4.0; Dr. Michael Pidwirny & Scott Jones via PhysicaGeography.net)

Figure 2.8.3 is a graph showing Earth’s energy balance as a function of latitude. The horizontal axis represents latitude from 90° north to 90° south, and the vertical axis shows energy in watts per square meter. Two curves are plotted: net incoming solar radiation and net outgoing longwave radiation. In low latitudes near the equator, incoming solar radiation exceeds outgoing radiation, creating an energy surplus. In high latitudes toward both poles, outgoing radiation exceeds incoming solar radiation, creating an energy deficit. The diagram highlights that excess heat from low latitudes is transported toward higher latitudes by the atmosphere and oceans.

The net heating imbalance between the equator and poles drives an atmospheric and oceanic circulation that climate scientists describe as a “heat engine.” In everyday use, we associate the word engine with automobiles, but for a scientist, an engine is any device or system that converts energy into motion. The climate is an engine that uses heat energy to sustain atmospheric and oceanic motion. Evaporation, convection, rainfall, winds, and ocean currents are all part of the Earth’s heat engine.

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# 2.9: Exercises

## Chapter 2: Exercises

1. Is the following statement about radiation true or false?

**Query 2.9.1**

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View this Interactive Content online



2. Describe the relationship between wavelength, energy emission, and the temperature of an object.
3. If the Earth's temperature 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:

**Query 2.9.2**

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# 3: AIR TEMPERATURE

## Introduction

Air Temperature is considered one of the most important factors in human comfort. It's often the largest number on the screen in weather apps, and many people even use "temperature" and "weather" interchangeably (that's incorrect, however, as there's much more to weather than just temperature). As described, air temperature is simply the temperature of the air. But how do we measure air temperature, and how is it influenced by various factors (such as latitude or elevation)? This chapter covers the fundamentals of air temperature, from its measurement to its seasonal changes.

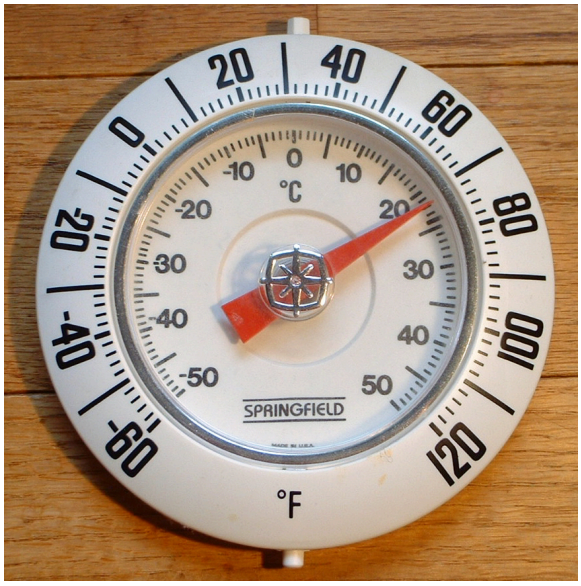



Figure 3.1: A common household dial thermometer. (Public Domain; Stilfehler, via Wikimedia Commons)

### Learning Objectives

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

- Determine if a weather station is correctly or incorrectly measuring air temperature.
- Identify some factors that wind and humidity have on perceived temperature.
- Determine if the air temperature should increase or decrease at a given time.
- Distinguish how each of the four factors influences a location's air temperature seasonally.

3.1: Air Temperature Measurements
3.2: Daily Temperature Variations
3.3: Seasonal Air Temperature Variations
3.4: Exercises for Chapter 3

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# 3.1: Air Temperature Measurements

## Thermometers

Temperature has been historically measured by a thermometer like the one shown in Figure 3.1.1. 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.

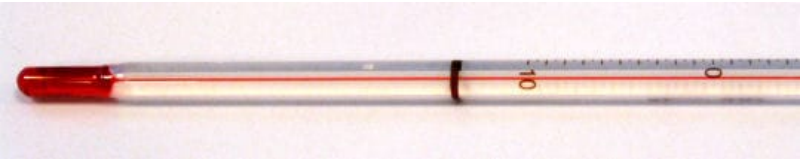


Figure 3.1.1: Liquid in glass thermometer (Public Domain; Chemical Engineer via Wikimedia Commons).

Another type of early thermometer records the minimum and maximum with mercury as shown in Figure 3.1.2. A marker is pushed upward (for the maximum) or downward (for the minimum).

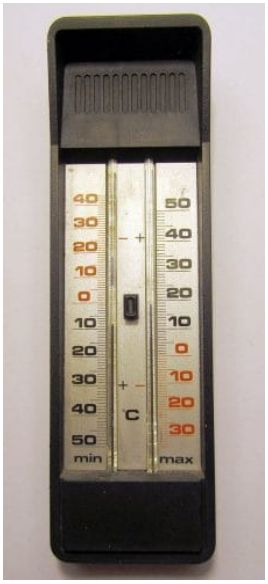



Figure 3.1.2: A thermometer device displaying temperature readings in degrees Celsius with separate scales for minimum and maximum temperatures. (CC-BY-SA 3.0; AtelierMonpli, via Wikimedia Commons).

Today, most thermometers are electronic. They use temperature-sensitive components, such as a temperature-sensitive resistor or capacitor, whose electrical current changes with temperature. Based on the instrument's electrical reading, temperature can be automatically recorded. An automated weather station using electronic thermometers is shown in Figure 3.1.3.



Figure 3.1.3: Electrical thermometers are used in systems such as this Elko Automated Surface Observing System (ASOS). This weather observation station is located in an open field and features multiple instruments and sensors mounted on poles to monitor atmospheric conditions. (CC-BY-SA 3.0; Famartin, via Wikimedia Commons).

Figure 3.1.4 shows a thermistor, which is commonly used in radiosondes. Radiosondes are attached to weather balloons for measuring temperature throughout the entire atmospheric column. We can observe in situ up to 50 km, halfway through the stratosphere, using a radiosonde balloon.



Figure 3.1.4: Radiosondes contain thermistor temperature sensors (CC-BY-SA 3.0; Sclereid, via Wikimedia Commons).

The World Meteorological Organization (WMO) has guidelines for the accurate placement of thermometers for recordkeeping purposes. The primary guides are that a thermometer must be placed:

- **1.5-2 meters above the ground**
- **in the shade**
- **in a white (or high albedo) ventilated shelter.**

Such a setup is shown in Figure 3.1.5:



Figure 3.1.5: An outdoor meteorological observation area containing multiple white, slatted instrument shelters mounted on legs to keep them above the ground. These shelters, known as Stevenson screens, are used to protect thermometers and other weather instruments from direct exposure while allowing air circulation. The site has a paved walkway, grassy surroundings, and additional weather equipment and towers in the background, indicating a professional weather station setup. (CC-BY 3.0; via Wikimedia Commons).

## Air Temperature Records

With over a century of standardized weather data collected across the world, we have been able to identify temperature extremes from across the world (along with other weather conditions).

The world's warmest recorded temperature was **134°F (56.7°C)**, recorded on July 10, 1913, at **Furnace Creek, in Death Valley, California, USA**.

The world's coldest recorded temperature was **-128.6°F (-89.2°C)**, recorded on July 21, 1983, at the **Vostok** research base in **Antarctica**.

## Air Temperature and Human Comfort

A temperature reading gives the air temperature, but wind and humidity can affect how the atmosphere feels to the human body. Figure 3.1.6 shows a wind chill graph where the wind speed and the temperature give an additional measure of temperature called the wind chill. When the air temperature is low and the wind is high, the temperature feels even colder and can cause frostbite within minutes (purple area).

The chart includes a grid where temperatures decrease across the top, and wind speeds increase down the side. Each cell shows the resulting wind chill value, illustrating how cold it feels under different combinations. Sections are shaded to represent frostbite risk times for 30, 10, and 5 minutes. A formula for calculating wind chill is shown in Figure 3.1.6, showing how air temperature and wind speed determine perceived temperature.

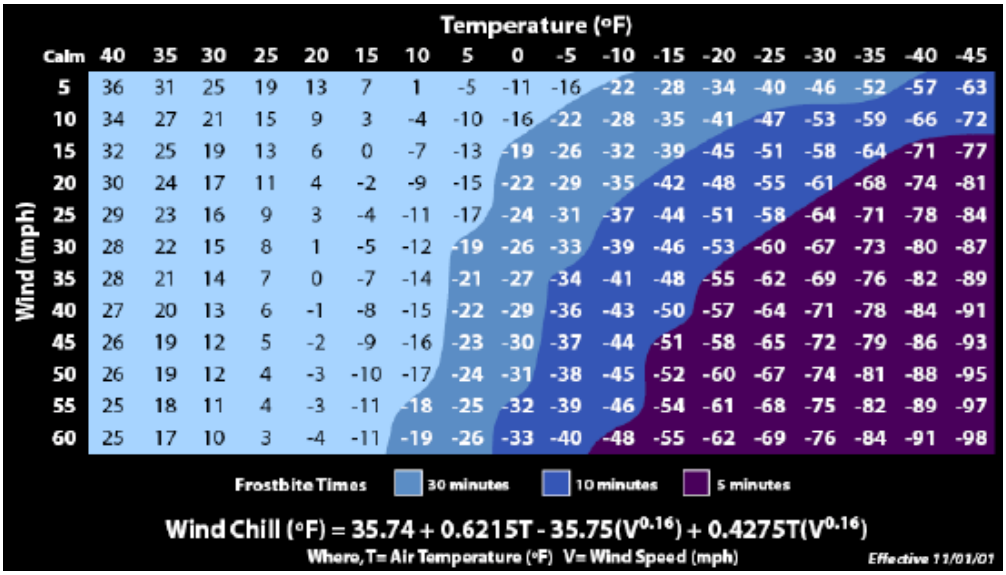



Figure 3.1.6: This chart shows wind chill values based on temperature in degrees Fahrenheit and wind speed in miles per hour. (Public Domain; via Wikimedia Commons).

The same table as in Figure 3.1.6 is shown in Table 3.1.1

*Table 3.1.1: Wind Chill Equivalent Temperatures (°F) by NWS*

| Wind speed (mph)↓ | Temperature (°F) | | | | | | | | | | | | | | | | | |
|---|---|---|---|---|---|---|---|---|---|---|---|---|---|---|---|---|---|---|
| Calm | 40 | 35 | 30 | 25 | 20 | 15 | 10 | 5 | 0 | -5 | -10 | -15 | -20 | -25 | -30 | -35 | -40 | -45 |
| 5 | 36 | 31 | 25 | 19 | 13 | 7 | 1 | -5 | -11 | -16 | -22 | -28 | -34 | -40 | -46 | -52 | -57 | -63 |
| 10 | 34 | 27 | 21 | 15 | 9 | 3 | -4 | -10 | -15 | -22 | -28 | -35 | -41 | -47 | -53 | -59 | -66 | -72 |
| 15 | 32 | 26 | 19 | 13 | 6 | 0 | -7 | -13 | -19 | -26 | -32 | -39 | -45 | -51 | -58 | -64 | -71 | -77 |
| 20 | 30 | 24 | 17 | 11 | 4 | -2 | -9 | -15 | -22 | -29 | -35 | -42 | -48 | -55 | -61 | -68 | -74 | -81 |
| 25 | 29 | 23 | 16 | 9 | 3 | -4 | -11 | -17 | -24 | -31 | -37 | -44 | -51 | -58 | -64 | -71 | -78 | -84 |
| 30 | 28 | 22 | 15 | 8 | 1 | -5 | -12 | -19 | -26 | -33 | -39 | -46 | -53 | -60 | -67 | -73 | -80 | -87 |
| 35 | 28 | 21 | 14 | 7 | 0 | -7 | -14 | -21 | -27 | -34 | -41 | -48 | -55 | -62 | -69 | -76 | -82 | -89 |
| 40 | 27 | 20 | 13 | 6 | -1 | -8 | -15 | -22 | -28 | -36 | -43 | -50 | -57 | -64 | -71 | -78 | -84 | -91 |
| 45 | 26 | 19 | 12 | 5 | -2 | -9 | -16 | -23 | -29 | -37 | -44 | -51 | -58 | -65 | -72 | -79 | -86 | -93 |
| 50 | 26 | 19 | 12 | 4 | -3 | -10 | -17 | -24 | -30 | -38 | -45 | -52 | -60 | -67 | -74 | -81 | -88 | -95 |
| 55 | 25 | 18 | 11 | 4 | -3 | -11 | -18 | -25 | -31 | -39 | -46 | -54 | -61 | -68 | -75 | -82 | -89 | -97 |
| 60 | 25 | 17 | 10 | 3 | -4 | -11 | -19 | -26 | -32 | -40 | -48 | -55 | -62 | -69 | -76 | -84 | -91 | -98 |

On the other hand, when the air is humid, it feels hotter. This is called the heat index, and we'll examine it further in the next chapter.

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# 3.2: Daily Temperature Variations

## Daily Highs and Lows

Daily changes are also called “diurnal.” The diurnal high and low temperatures of a location depend on many factors, but most importantly on the balance between incoming solar radiation and outgoing long-wave radiation. Remember that, in addition to the sun emitting solar radiation, Earth also emits infrared radiation.

## Daytime Temperatures

Starting at sunrise, the sun rises and incoming solar radiation increases. To understand why this happens, you can shine a flashlight on a surface. When the flashlight is held at a low angle, the illuminated area of the surface is large but dim (Figure 3.2.1). The same is true when the sun is low on the horizon; solar intensity is low. On the other hand, as you increase the angle of the flashlight until it is pointing straight down onto the surface, the illuminated surface becomes smaller, but brighter (Figure 3.2.2). The same is true when the sun is high in the sky: solar intensity increases as it rises.

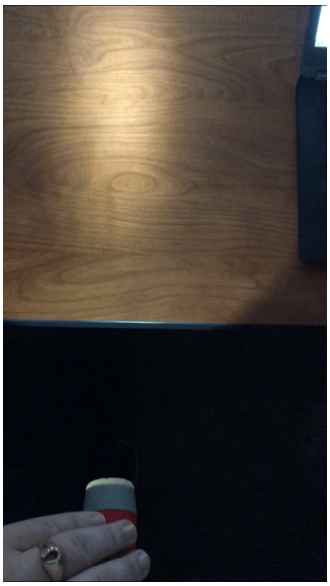


Figure 3.2.1: When a flashlight is shining on a surface from a low angle, the illuminated surface is large, but dim. (Public Domain, Alicia Mullens)



Figure 3.2.2: When the flashlight is shining directly overhead, the illuminated surface is smaller, but significantly brighter. (Public Domain, Alicia Mullens)

Once the sun rises, incoming solar radiation quickly exceeds outgoing long-wave radiation, and Earth begins to warm. As solar energy strikes the Earth's surface each morning, a shallow 1–3-centimetre (0.39–1.18 in) layer of air directly above the ground is heated by conduction. Heat exchange between this shallow layer of warm air and the cooler air above is very inefficient. On a warm summer's day, for example, air temperatures may vary by 16.5 °C (30 °F) from just above the ground to chest height. As the morning progresses, the amount of incoming solar radiation increases, further warming air temperatures. By noon, the sun peaks in the sky, but incoming solar radiation is still significantly more than the outgoing long-wave radiation Earth is emitting, and so Earth continues to warm. As the ground warms, the amount of long-wave radiation it emits increases. In the afternoon, the sun begins to lower in the sky, and the amount of incoming solar radiation declines. However, because the Earth is still warming, the amount of outgoing long-wave radiation continues to increase. Eventually, the two (incoming solar radiation and outgoing long-wave radiation) equal each other. At this time, warming stops, and the highest daytime temperature is achieved. While other factors (advection, cloud cover, etc.) can affect the timing of the highest daily temperature, the typical time of the highest temperature is between **3 and 5 pm**. The relationship between incoming and outgoing radiation and how they affect temperature is shown in Figure 3.2.3.

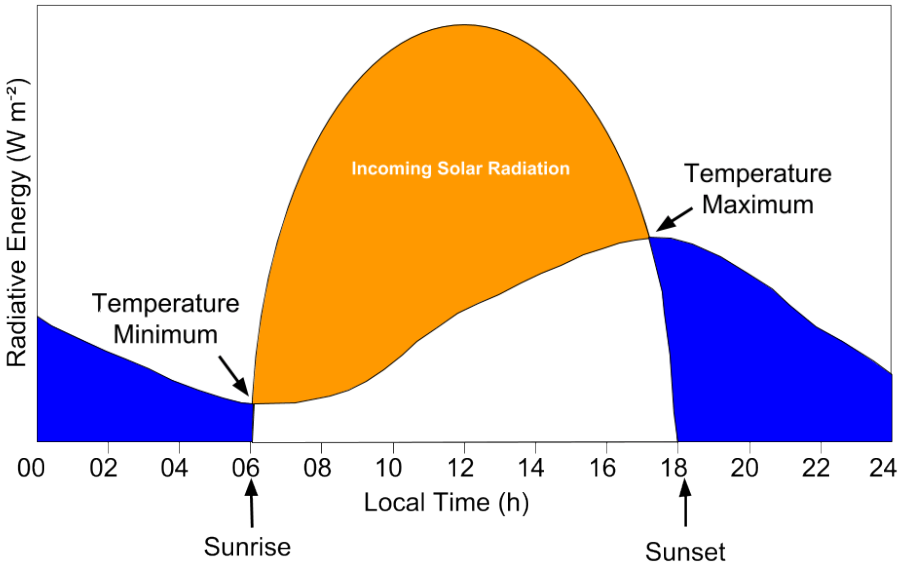



Figure 3.2.3: The amount of radiative energy (y-axis, in W m-2) as a function of time (x-axis, in hours). The shaded area between the minimum and maximum temperature represents net warming, and the shaded areas before the minimum and after the maximum represent net cooling. The temperature is coolest at sunrise, warms during the day, and reaches its maximum temperature in the late afternoon. (CC BY-SA 4.0; Britt Seifert via Atmospheric Processes and Phenomena).

Temperature lag, also known as thermal inertia, is an important factor in diurnal temperature variation. The peak daily temperature generally occurs *after* noon, as the air continues to absorb net heat from morning through noon and for some time thereafter. After the daytime high temperature is reached, the amount of incoming solar radiation continues to decrease as the sun sets, and Earth begins to cool. This is why, even though the sun is still out in the evening, it may feel significantly cooler than earlier in the day.

## Nighttime Temperatures

At night, no sunlight enters, yet Earth continues to emit long-wave radiation. As a result, Earth loses heat and begins to cool. This cooling process is called **radiational cooling**. This continues throughout the night as Earth loses radiation without any incoming sunlight. As a result, the coldest temperature of the night is typically **at or just after sunrise**. Because air is a poor conductor of heat, radiational cooling is most extreme on surfaces. As a result, **temperatures near the ground (and on surfaces) can be significantly cooler than those in the air just a few feet above**. You can experience this when walking outside barefoot at night: the ground feels considerably cooler than the air at eye level. Another consequence of this is the formation of dew on surfaces, even if the air above those surfaces is not cold enough to become saturated. This will be covered in a future chapter.

While radiational cooling occurs every night, certain conditions can enhance it, resulting in extremely cold temperatures on some nights. This happens when nights are:

- **Long**, which allows for more extended radiational cooling.
- **Clear**, which allows long-wave radiation to escape into space.
- **Dry**, because moisture is an excellent reservoir of heat, the absence of moisture allows for additional cooling.
- **Calm**, because winds mix in warmer air from above with cooler air near the ground, the lack of winds prevents mixing at the surface.

One final interesting feature of nighttime temperatures is that they can be significantly cooler on valley floors than on the adjacent hillsides. This occurs because cooler air sinks into the valley floor, leaving warmer air on the hillsides. Because many agricultural regions, such as central California, lie on valley floors, farmers use various techniques, including orchard heaters, to prevent their crops from freezing.

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# 3.3: Seasonal Air Temperature Variations

## Seasonal Variations

Just as air temperatures vary throughout the day, they also vary seasonally. The discussion below focuses on daily average high temperatures and their seasonal variation. Still, you could apply these same ideas to overnight low temperatures as well and get the same general results. A location's seasonal air temperature range can be determined by subtracting the coldest average annual temperature from the warmest average annual temperature. For example, San Jose, California, has the warmest average temperature in August at 82°F and in January at 60°F. yielding a 22°F range (Figure 3.3.1, and Table 3.3.1).

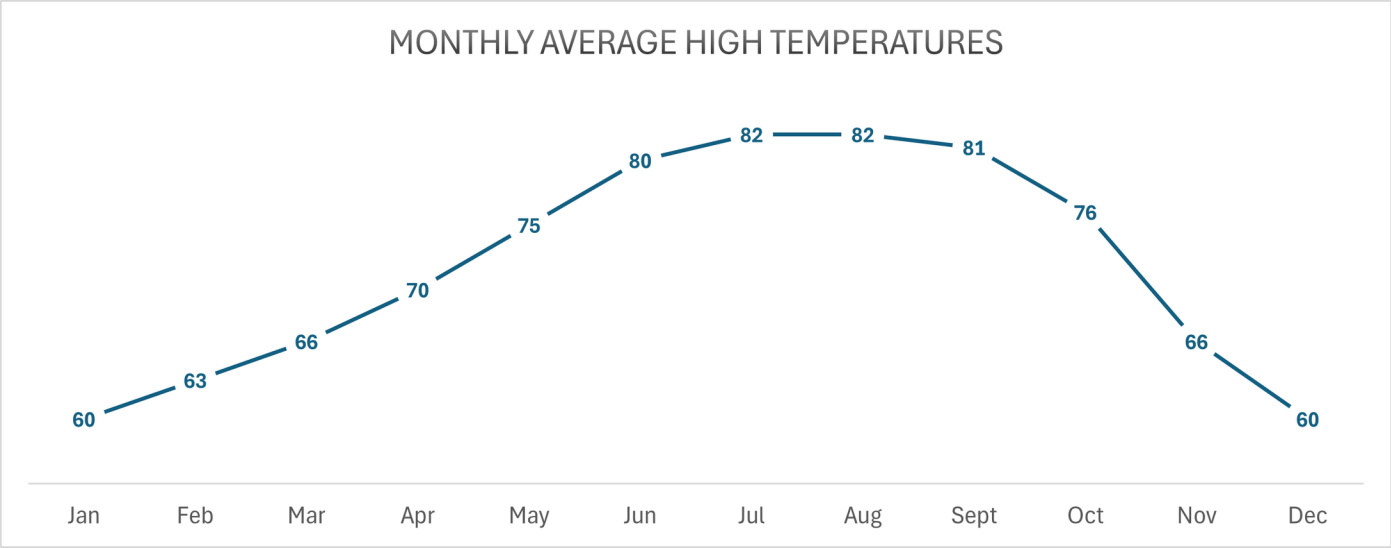



Figure 3.3.1: Average monthly high temperatures for San Jose, California. (CC-BY-SA 4.0; Alicia Mullens)

*Table 3.1.1: Average monthly high temperatures for San Jose, California.*

| Month | Jan | Feb | Mar | Apr | May | Jun | Jul | Aug | Sept | Oct | Nov | Dec |
|---|---|---|---|---|---|---|---|---|---|---|---|---|
| Temperature (°F) | 60 | 63 | 66 | 70 | 75 | 80 | 82 | 82 | 81 | 76 | 66 | 60 |

A location's seasonal temperature variation depends on four main factors, often referred to as the "controls" on seasonal air temperature variations:

1. Latitude
2. Proximity to Ocean vs. Land
3. Elevation
4. Ocean Currents

Let's investigate each of these a little further:

## Latitude

The latitude of a location has two main effects on its temperature. The first one is on the intensity of the incoming sunlight. From the previous section, recall that when the sun is higher in the sky, the intensity of incoming solar radiation is higher. As a result, locations closer to the equator see warm temperatures, while locations further from the equator see colder temperatures. The other important impact comes from the length of the day: locations near the equator experience nearly constant day length year-round, while locations further from the equator see greater and greater variation. In fact, locations poleward of 66.5°N/S have at least one full 24-hour day of light and one full 24-hour day of darkness per year, with the frequency rapidly increasing until you get to the pole, which sees six consecutive months of daylight followed by six straight months of darkness. Putting these two

concepts together, the main takeaway from the discussion of latitude's impact on temperatures is that locations near the equator experience more consistent temperatures, while those farther from the equator experience greater temperature variation. This is illustrated in Figure 3.3.2 and Table 3.3.2, which shows the average monthly high temperatures for Honolulu (latitude 21°N) vs. Anchorage (latitude 61°N).

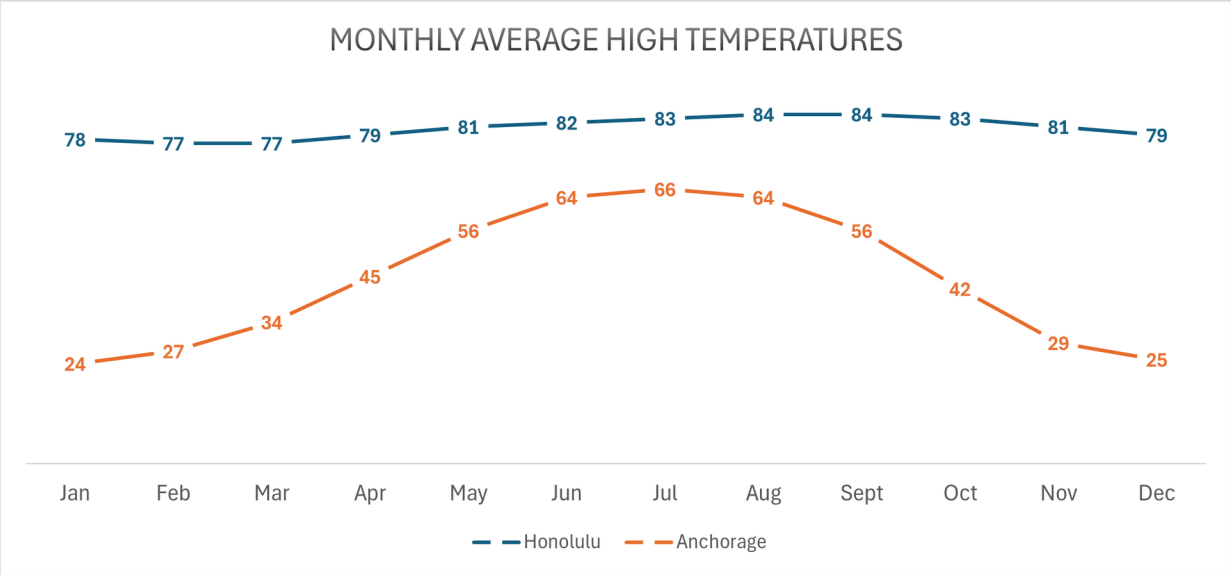



Figure 3.3.2: Average monthly high temperatures for Honolulu, Hawaii, and Anchorage, Alaska. Honolulu's temperatures remain steady between about 77 and 84 degrees year-round. Anchorage's temperatures start near 24 degrees in January, rise to about 66 degrees in July, then drop back to around 25 degrees in December. (CC-BY-SA 4.0; Alicia Mullens)

*Table 3.3.2: Average monthly high temperatures for Honolulu, Hawaii, and Anchorage, Alaska.*

| Month | Jan | Feb | Mar | Apr | May | Jun | Jul | Aug | Sept | Oct | Nov | Dec |
|---|---|---|---|---|---|---|---|---|---|---|---|---|
| Honolulu Temperature (°F) | 78 | 77 | 77 | 79 | 81 | 82 | 83 | 84 | 84 | 83 | 81 | 79 |
| Anchorage Temperature (°F) | 24 | 27 | 34 | 45 | 56 | 64 | 66 | 64 | 56 | 42 | 29 | 25 |

Honolulu, which is closer to the equator, experiences minimal temperature variation (only 7°F between summer highs and winter lows). In contrast, Anchorage, which is closer to the north pole, sees significantly more (a difference of 42°F between summer highs and winter lows). It should be noted that both cities are coastal cities located at/near sea level. As you will see below, both factors do indeed influence seasonal temperature variations, but in this case, the primary difference between Honolulu and Anchorage is latitude.

## Proximity to Ocean vs. Land

Have you ever been to the beach on a bright, warm day and tried to walk on the sand barefoot? It's very likely that the sand you were walking on felt very hot to your feet, and that probably left you running for the cooler ocean water. That dichotomy (warmer sand vs. cooler ocean water) has nothing to do with one receiving more/less heat than the other; assuming they are right next to each other, both are exposed to the same amount of sunlight. The big difference between the two is **specific heat**, the amount of heat required to raise an object's temperature by 1°C. Water has a specific heat that is, on average, five times higher than that of most land cover (soil, concrete, sand, etc.), and as a result, water requires substantially more heat to increase its temperature than land does. Consequently, water does not warm up or cool down as much as land does throughout the day AND throughout the year. During summertime, ocean waters remain relatively cool, providing a pleasant respite from hot coastal temperatures, while in wintertime, ocean waters remain relatively warm, preventing coastal regions from experiencing the biting cold common inland. The general rule is that coastal areas experience less seasonal temperature variation than inland areas. An excellent example of this is shown in Figure 3.3.3 and Table 3.3.3, which compares the monthly average high temperatures in Long Beach, California, and Dallas, Texas. Both cities have similar latitudes and elevations, but Long Beach is near the coast while Dallas is far inland.

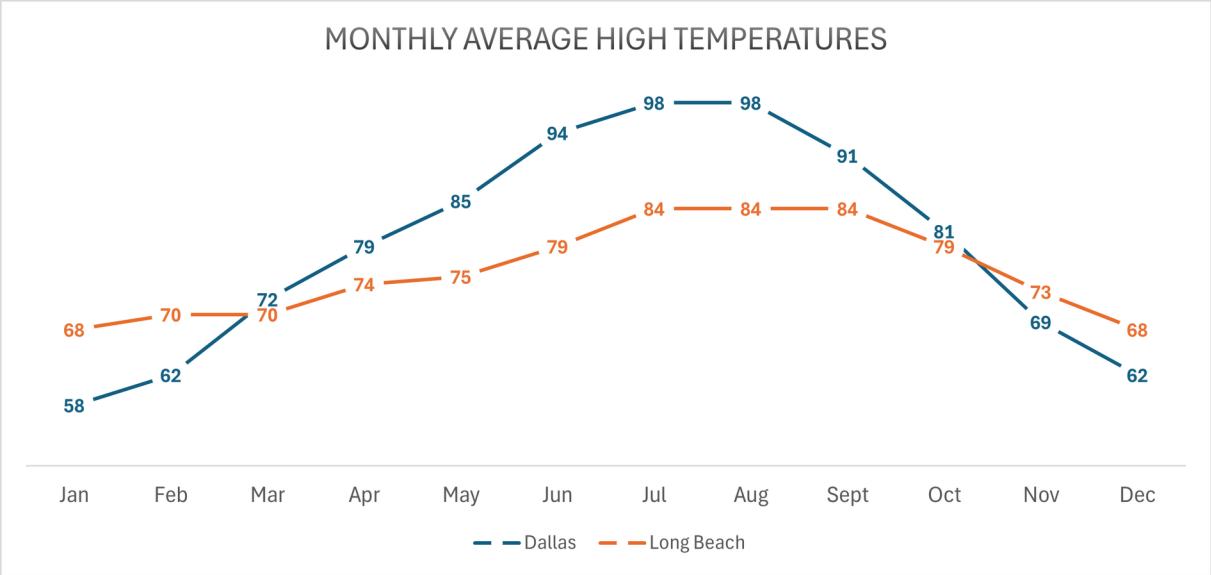



Figure 3.3.3: Average monthly high temperatures for Long Beach, California vs. Dallas, Texas. Dallas starts around 58 degrees in January, rises to about 98 degrees in July and August, then drops back to 62 degrees in December. Long Beach begins near 68 degrees in January, peaks around 84 degrees from July through September, and returns to 68 degrees in December. (CC-BY-SA 4.0; Alicia Mullens)

*Table 3.3.3: Average monthly high temperatures, Long Beach, California, and Dallas, Texas.*

| Month | Jan | Feb | Mar | Apr | May | Jun | Jul | Aug | Sept | Oct | Nov | Dec |
|---|---|---|---|---|---|---|---|---|---|---|---|---|
| Long Beach Temperature (°F) | 68 | 70 | 70 | 74 | 75 | 79 | 84 | 84 | 84 | 49 | 73 | 68 |
| Dallas Temperature (°F) | 58 | 62 | 72 | 79 | 85 | 94 | 98 | 98 | 91 | 81 | 69 | 62 |

Long Beach stays relatively mild year-round (with average high temperatures of 68°F in winter and 84°F in summer). At the same time, Dallas experiences significantly more variation (with average high temperatures of 58°F in winter and 98°F in summer). These averages only tell part of the story, too; winter weather in Dallas can be notorious for icy conditions, biting cold, and even blizzards, while Long Beach experiences nothing of the sort. Summer weather in Dallas isn't much better, with weeks of temperatures above 100°F, while Long Beach may experience a few days (if that) a year with such extreme heat. Their respective proximity to the ocean vs. the land is the main reason for such differences.

## Elevation

Suppose you live in a tropical location, such as Hawaii, or a relatively mild climate, such as much of California. At sea level, in both climates, snow is rare, if not nonexistent (especially in Hawaii), but that doesn't mean either state never experiences it. In fact, the snowy weather typical of winter is common at higher elevations in both states. In Hawaii, snow is mainly restricted to the highest peaks of the Big Island's volcanic mountains (such as Mauna Loa), while in California, snow is common at higher elevations in the Sierra and Cascade mountains (and to a lesser extent in the transverse ranges of Southern California). This brings us to our third main factor on seasonal temperature variations: Elevation. All else being equal (latitude, proximity to the ocean, etc.), a location at a higher elevation will be cooler year-round than one closer to sea level. Figure 3.3.4 and Table 3.3.4 compare the average monthly high temperatures for Sacramento to South Lake Tahoe, California. Both cities are inland and at nearly the same latitude, but Sacramento has an average elevation of about 100 feet, while South Lake Tahoe has an average elevation of about 6,200 feet.

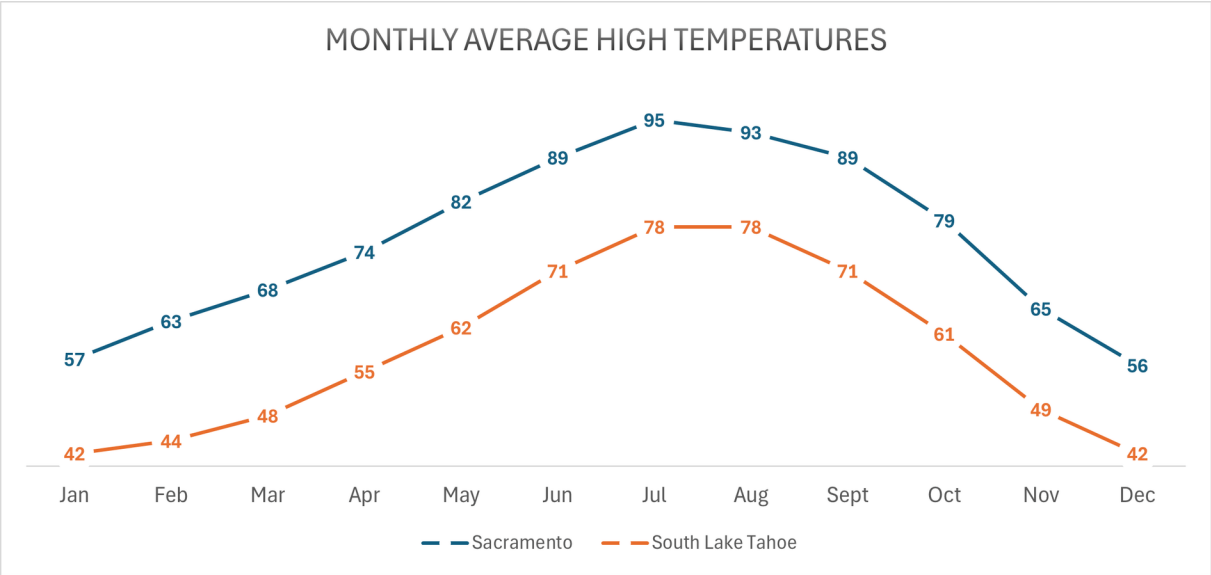



Figure 3.3.4: Average monthly high temperatures for Sacramento vs. South Lake Tahoe. Sacramento's temperatures start at around 57 degrees in January, rise to about 95 degrees in July, then fall to 56 degrees in December. South Lake Tahoe's temperatures begin near 42 degrees in January, peak around 78 degrees in July and August, and drop back to 42 degrees in December. (CC-BY-SA 4.0; Alicia Mullens)

*Table 3.3.4: Average monthly high temperatures for Sacramento, California, and South Lake Tahoe, California.*

| Month | Jan | Feb | Mar | Apr | May | Jun | Jul | Aug | Sept | Oct | Nov | Dec |
|---|---|---|---|---|---|---|---|---|---|---|---|---|
| Sacramento Temperature (°F) | 57 | 63 | 68 | 74 | 82 | 89 | 95 | 93 | 89 | 79 | 65 | 56 |
| South Lake Tahoe Temperature (°F) | 42 | 44 | 48 | 55 | 62 | 71 | 78 | 78 | 71 | 61 | 49 | 42 |

If you notice, both cities have very similar seasonal temperature variations (38°F in Sacramento and 36°F in South Lake Tahoe), but South Lake Tahoe is, on average, about 15-20°F cooler than Sacramento. This difference is purely due to their elevation differences: the lower-elevation Sacramento is warmer year-round, while the higher-elevation South Lake Tahoe is cooler year-round.

## Ocean Currents

The final factor that controls a location's seasonal temperature variation is the ocean current it is next to. This factor obviously has less influence the further inland you go, but it can have a substantial impact on locations near the coast. While the details of ocean currents will be saved for a later chapter, this discussion will consider two types of ocean currents: warm ocean currents that originate near the equator and carry warm ocean waters towards the poles, and cold ocean currents that originate closer to the poles and carry cold ocean waters towards the equator. These ocean currents are an essential component of Earth's energy budget and play a key role in balancing Earth's temperature differences. Cold ocean currents are generally found along the west coasts of continents (such as the California Current along the west coast of North America).

In contrast, warm ocean currents are usually found along the east coasts of continents (such as the Gulf Stream along the east coast of North America). In the United States, the West Coast is located next to the cold California Current, while the East Coast is located next to the warm Gulf Stream. Hawaii, in the central Pacific, is still subject to the eastward-flowing equatorial current. Still, the temperature differences caused by this current are generally less significant than the temperature differences caused by the warm and cold currents that parallel extensive coastlines.

Figure 3.3.5 and Table 3.3.5 show the influence of ocean currents on two coastal cities with similar latitudes and elevations. San Diego, California, is located next to the cooler California current, while Savannah, Georgia, is located next to the warmer Gulf Stream. It should be noted that, for a reason that will be discussed in a later chapter (specifically, both locations are dominated by mid-latitude westerlies that blow from west to east in the wintertime), the differences between San Diego and Savannah are more noticeable in summer than in winter.

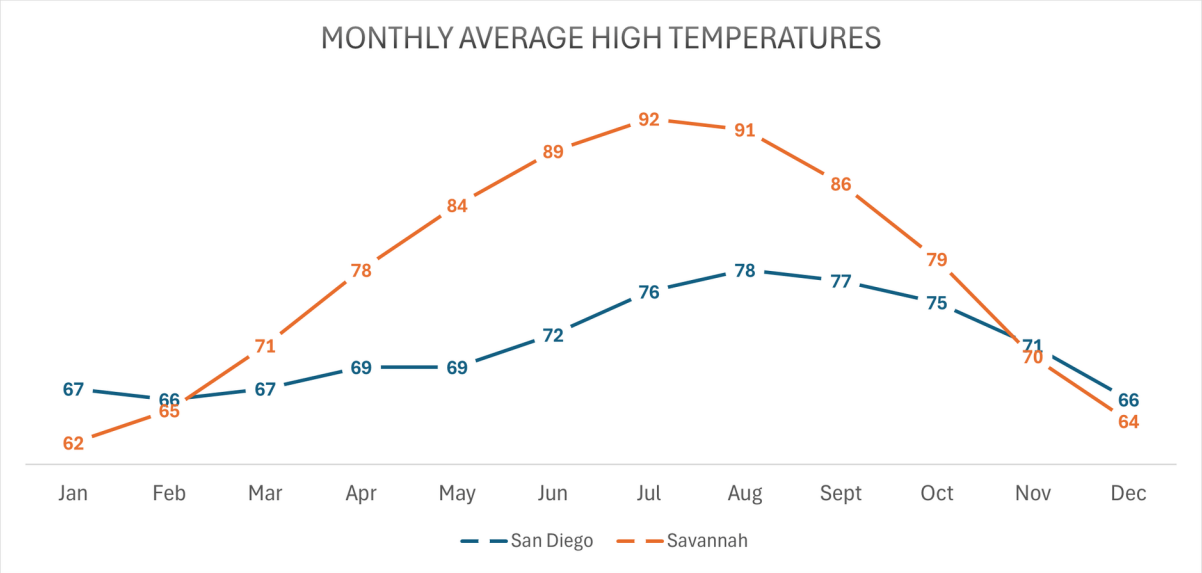



Figure 3.3.5: San Diego's temperatures stay relatively steady, starting around 67 degrees in January, peaking near 78 degrees in August, and ending at 66 degrees in December. Savannah's temperatures start around 62 degrees in January, rise to about 92 degrees in July, then decline to 64 degrees in December. (CC-BY-SA 4.0; Alicia Mullens)

*Table 3.3.5: Average monthly high temperatures for San Diego, California, and Savannah, Georgia.*

| Month | Jan | Feb | Mar | Apr | May | Jun | Jul | Aug | Sept | Oct | Nov | Dec |
|---|---|---|---|---|---|---|---|---|---|---|---|---|
| San Diego Temperature (°F) | 67 | 66 | 67 | 69 | 69 | 72 | 76 | 78 | 77 | 75 | 71 | 66 |
| Savannah Temperature (°F) | 62 | 65 | 71 | 78 | 84 | 89 | 92 | 91 | 86 | 79 | 70 | 64 |

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# 3.4: Exercises for Chapter 3

## Chapter 3: Exercises

1. Consider the following thermometer set-ups. Which ones would correctly measure temperature, according to WMO guidelines?
   a. Your car's thermometer.
   b. An airport weather station thermometer, placed 1.75 meters above the ground in a white, ventilated shelter.
   c. You are holding a glass thermometer while gripping the bulb.
   d. A sheltered/ventilated thermometer on the top of a 2-meter-tall buoy.
2. Why is the warmest time of day in the afternoon rather than at noon (when the sun is highest in the sky)?
3. Why do clear nights feel so cold when compared to cloudy nights?
4. Consider the following locations. Which one is most likely to have the most significant seasonal temperature variation, and why (consider their geography, latitude, elevation. It's okay to look up information about their location)?
   a. Chicago, Illinois
   b. Eureka, California
   c. Denver, Colorado
   d. Hilo, Hawaii
   e. Miami, Florida

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# 4: WATER VAPOR

## Introduction

Water can exist as a solid, liquid, or gas under typical Earth conditions. Figure 4.1 illustrates the water cycle on Earth. As we learned earlier, the process by which liquid water becomes water vapor is called **evaporation**, and it requires energy. As the Sun heats the Earth, liquid water molecules from oceans, lakes, and rivers absorb energy and change into water vapor. Water vapor is also released by plants through a process called **transpiration**. The opposite process is called **condensation**, where water vapor cools at higher altitudes and becomes liquid water or ice, releasing energy. Tiny droplets condense on dust or sand particles in the air. Condensation is significant in atmospheric science because it is the process that allows clouds to form.

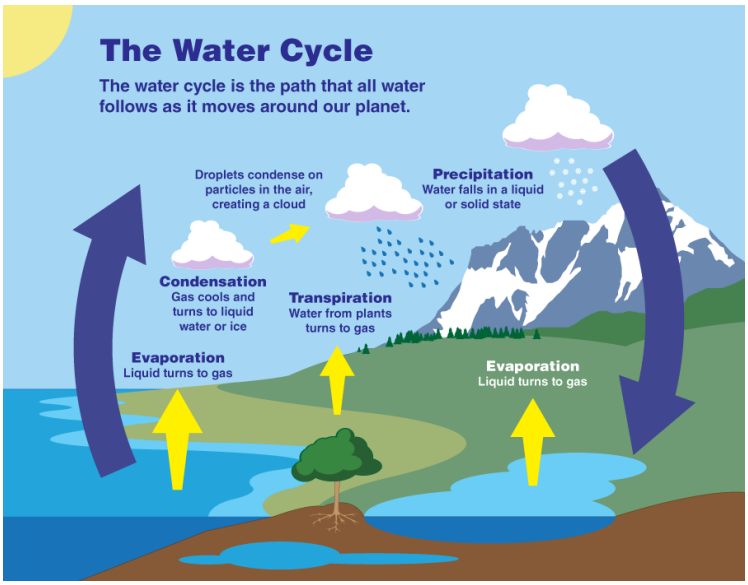



Figure 4.1: Earth's water cycle is the path that water takes as it moves around Earth. This includes times when it flows, such as during rainfall, and places where it's stored, such as in ice, groundwater, lakes, rivers, and the ocean. Liquid water from such reservoirs evaporates, condenses into clouds, and returns to the surface as precipitation. (Public Domain; The Water Cycle via NASA/JPL-Caltech).

Clouds are composed of millions and billions of tiny liquid water droplets. How do they form? Why are they there? Before we can understand clouds in the atmosphere, we need to explore concepts like how humidity is defined and what saturation means.

### Learning Objectives

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

- Differentiate between relative humidity, specific humidity, absolute humidity, wet-bulb temperature, mixing ratio, and dew point.
- Use the Heat Index to determine personal comfort.
- Describe the conditions for saturation to occur.
- Distinguish between Dew and Frost.
- Describe how fog forms and the different types of fogs.
- Distinguish between different types of clouds.

4.1: Saturation

4.2: Humidity

4.3: The Heat Index

4.4: Dew and Frost
4.5: Fog
4.6: Clouds
4.7: Exercises

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# 4.1: Saturation

## Evaporation and Condensation

Imagine a beaker filled halfway with water as shown in Figure 4.1.1. Let's put a lid on it to constrain water molecule movement and eliminate the wind's influence on evaporation. At the water surface, some molecules absorb thermal energy from the environment, undergo phase change, and enter the air as water vapor (evaporation). These molecules move about suspended in the air, but they have nowhere to go because the top is closed. Thus, a fraction of the molecules near the surface absorb onto the surface and recombine to form liquid water (condensation). As evaporation increases, the water level in the beaker decreases, as more water molecules enter the closed air above the liquid. This occurs because, at typical room temperature conditions, the number of molecules evaporating exceeds the number of water molecules condensing back onto the surface. This is called net evaporation. As the number of water vapor molecules in the air increases, so does the number of vapor molecules condensing back to the liquid phase. After some time, the amount of water entering the air from evaporation equals that condensing. When this occurs, the air is said to be **Saturated.**

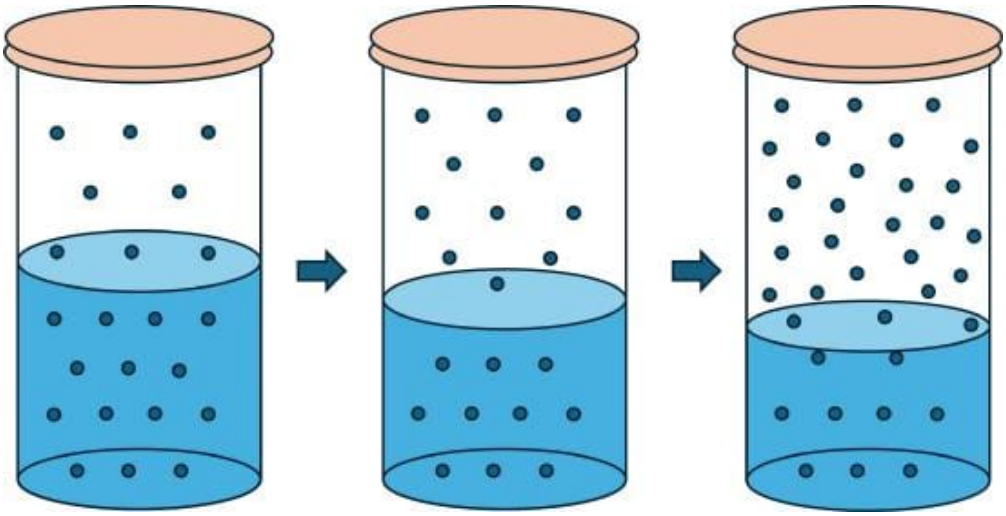


Figure 4.1.1: Evaporation and condensation in an enclosed beaker of water. (CC BY 4.0; Neel Desai)

The air's saturation level is directly related to its temperature. As air temperature increases, air becomes thinner, and more water vapor molecules have more energy to remain in the gas phase. As the temperature decreases, air becomes denser, and air and water molecules slow down, increasing the likelihood that water vapor will condense onto surfaces. Thus, at higher temperatures, there is more water vapor in the air than at lower temperatures. Saturation occurs when air contains the maximum amount of water vapor possible for its given temperature. When that happens, the rate of condensation equals the rate of evaporation.

## Vapor pressure at saturation

Every gas in the atmosphere exerts pressure; for example, vapor pressure contributes to the total atmospheric pressure. Since the total air pressure near sea level is nearly 1000 hPa, nitrogen (78%) exerts nearly 780 hPa, and oxygen (21%) exerts nearly 210 hPa. Specifically, for water vapor, the more it is added to the atmosphere, the higher the vapor pressure. The units for vapor pressure are the same as those for pressure and can be expressed in pascals (Pa), hectopascals (hPa), or kilopascals (kPa).

The amount of water vapor that the atmosphere can contain depends on temperature. As noted earlier, lower-temperature air contains less water vapor than higher-temperature air. If we consider this quantitatively in terms of pressure, the *saturation vapor pressure* is the pressure exerted by water vapor molecules moving over a liquid water surface. When the partial pressure of water vapor equals the saturation vapor pressure, the air is saturated with water vapor. Figure 4.1.2 shows the relationship between the saturation vapor pressure and temperature in the atmosphere. 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 the saturation of water vapor.

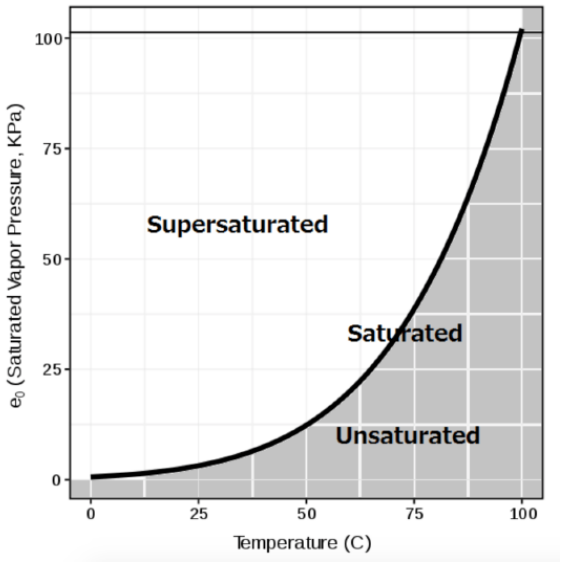



Figure 4.1.2: A graph of the saturation vapor pressure as a function of temperature showing the exponential relationship between the two from the Clausius-Clapeyron equation. (CC BY-SA 4.0; Modified from Firepulaski via Wikimedia Commons).

In Figure 4.1.2, notice the saturation vapor pressure value at the boiling temperature, 100 °C. The saturation vapor pressure is 101.325 kPa (1013.25 hPa), which is approximately equal to 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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# 4.2: Humidity

## Absolute Humidity

Humidity measures the amount of water vapor in the air. The amount of water vapor in the air depends on the controls over evaporation discussed earlier. A meteorologist can express air humidity in several ways. Each humidity measurement is, to some degree, controlled by air temperature.

Absolute humidity is the weight of water vapor per unit volume of air, usually measured in units of grams of water vapor per cubic meter of air ($g/m^3$). Absolute humidity is not often used to express the moisture content of air because it is sensitive to changes in both air temperature and atmospheric pressure. For instance, let's say a 1-cubic-meter parcel of air at the surface contains 2 grams of water. This gives us an Absolute Humidity of 2 $g/m^3$. Now lift the parcel of air upwards into the atmosphere. As the air rises, the decrease in atmospheric pressure on the parcel allows it to expand and occupy more space. Let's say the parcel doubles in size due to uplift. Before rising, the absolute humidity was 2 $g/m^3$. As the air doubles in volume, the new absolute humidity is 1 $g/m^3$. In actuality, the parcel still contains the same amount of water, 2 grams. But given how absolute humidity is calculated, it appears that the amount of water in the air has decreased. Thus, instead of absolute humidity, we use a measure that is not sensitive to changes in air volume.

## Specific Humidity

Specific humidity is the mass of water vapor in the air per unit mass of air, including the mass of water vapor. The units of measurement are grams of water vapor per kilogram of air (g/kg). Because weight is not significantly affected by temperature or atmospheric pressure, specific humidity is a much more useful measure of humidity. Another measure very similar to specific humidity is the mixing ratio. The **mixing ratio** is the weight of water vapor per unit weight of dry air. Because the atmosphere contains so little moisture by volume, the mixing ratio is virtually equal to the specific humidity.

## Relative Humidity

Relative humidity is the most commonly used way of describing atmospheric moisture. It is the ratio of the amount of water vapor present in the air to the maximum amount needed for saturation at a given pressure and temperature. It is typically multiplied by 100 and expressed as a percent.

$$RH = \frac{water\ vapor\ content}{water\ vapor\ capacity} \times 100\% \tag{4.2.1}$$

Relative humidity indicates how close the air is to saturation, not how much water vapor it contains. Regardless, we know that the air's water vapor saturation level depends on its temperature. We know that as air temperature increases, its water vapor capacity increases. That is, as the air temperature increases, it can keep more water in the vapor state. Relative humidity can also be expressed as a ratio of the actual vapor pressure to the saturation vapor pressure.

$$RH = \frac{actual\ vapor\ pressure}{saturation\ vapor\ pressure} \times 100\% \tag{4.2.2}$$

Thus, air at 50 percent relative humidity contains half the amount of water vapor needed to saturate it. Air with 100 percent relative humidity is said to be saturated. Interestingly, relative humidity can exceed 100 percent. When that happens, air is said to be **supersaturated**.

## Dew Point Temperature

The dew point is the temperature the air must be cooled to (at constant pressure) to reach 100 percent relative humidity, or, in other words, saturation. At this point, the air cannot hold any more water in its gaseous form. If the air were cooled even further, water vapor would condense into liquid water, typically as fog or precipitation. The dew point temperature is a good indicator of

the amount of moisture in the air. High dew points indicate higher water vapor content, while low dew points indicate low water vapor content in the air. When the air temperature matches the dew point, the air is saturated, and the relative humidity equals 100 percent.

It is called the "dew point" because dew forms on surfaces when the air reaches saturation. When the dew point is below the freezing point of water, it is also called the *frost point*.

## Wet-Bulb Temperature

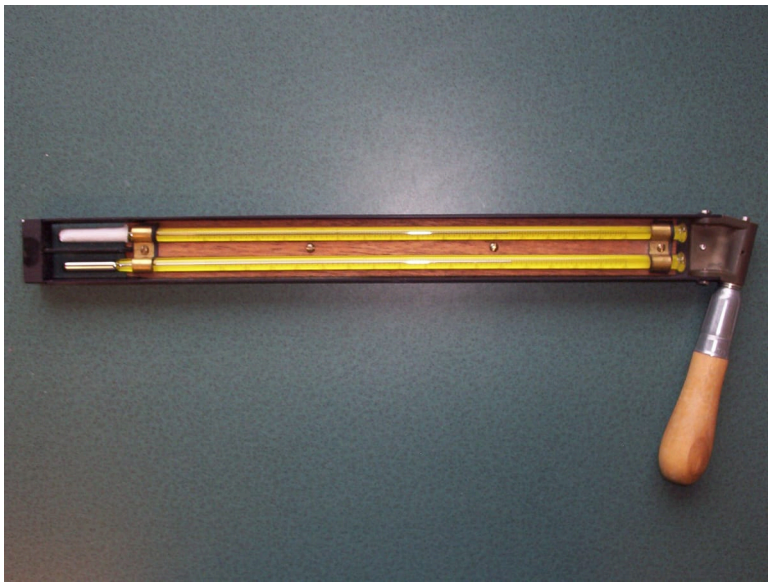


Figure 4.2.1: A sling psychrometer consists of two thermometers. A small piece of cotton or fabric is wetted with distilled water and placed over one of the thermometer's bulbs. A handle on the other end allows the device to be whirled in the air until the wet bulb is completely dry. (Public Domain; CambridgeBayWeather via Wikimedia Commons)

Wet-bulb temperature is the lowest temperature that can be achieved if water evaporates within the air. In other words, it is the temperature the air needs to be cooled to for it to saturate. It isn't easy to calculate, but it is easy to measure. To measure the wet-bulb temperature, simply wrap a wet cloth around the thermometer's bulb. Typically, this thermometer is attached to an apparatus called a sling psychrometer, as shown in Figure 4.2.2. The handle at the end makes it easy to whirl it through the air, creating a lot of airflow over the thermometer's wet cloth. Evaporation from the wet cloth cools the temperature measured by one of the thermometers. The temperature measured by the dry bulb is called the *dry-bulb* temperature or current air temperature. The temperature measured by the wet bulb after evaporative cooling is called the *wet-bulb* temperature. The wet-bulb temperature is always lower than the air temperature (or dry-bulb temperature) when relative humidity is less than 100%. When the relative humidity is 100%, the wet-bulb temperature equals the air temperature.

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# Heat Index Temperature

"It's not the heat, it's the humidity". That's a common phrase you may have heard in the summer, but it's actually both. The heat index, also known as apparent temperature, is the temperature a person feels when relative humidity is combined with air temperature. This has important implications for human comfort. When the body gets too hot, it sweats to cool itself. If perspiration cannot evaporate, the body cannot regulate its temperature. Evaporation is a cooling process. When perspiration evaporates from the body, it effectively lowers body temperature. When the atmospheric moisture content (i.e., relative humidity) is high, the rate of evaporation from the body decreases. In other words, the human body feels warmer in humid conditions. The opposite is true when relative humidity decreases, as the rate of perspiration increases. The body actually feels cooler in arid conditions. There is a direct relationship between the air temperature and relative humidity and the heat index, as shown in Table 4.3.1. As air temperature and relative humidity rise, the heat index rises.

*Table 4.3.1: Heat Index Temperatures (°F) for various relative humidity values by NOAA. The higher the Heat Index, the greater the risk of Heat Stroke and Heat Exhaustion.*

| Relative Humidity (%) ↓ Temp (°F) → | 80 | 82 | 84 | 86 | 88 | 90 | 92 | 94 | 96 | 98 | 100 | 102 | 104 | 106 | 108 | 110 |
|---|---|---|---|---|---|---|---|---|---|---|---|---|---|---|---|---|
| 40 | 80 | 81 | 83 | 85 | 88 | 91 | 94 | 97 | 101 | 105 | 109 | 114 | 119 | 124 | 130 | 136 |
| 45 | 80 | 82 | 84 | 87 | 89 | 93 | 96 | 100 | 104 | 109 | 114 | 119 | 124 | 130 | 137 | |
| 50 | 81 | 83 | 85 | 88 | 91 | 95 | 99 | 103 | 108 | 113 | 118 | 124 | 131 | 137 | | |
| 55 | 81 | 84 | 86 | 89 | 93 | 97 | 101 | 106 | 112 | 117 | 124 | 130 | 137 | | | |
| 60 | 82 | 84 | 88 | 91 | 95 | 100 | 105 | 110 | 116 | 123 | 129 | 137 | | | | |
| 65 | 82 | 85 | 89 | 93 | 98 | 103 | 108 | 114 | 121 | 128 | 136 | | | | | |
| 70 | 83 | 86 | 90 | 95 | 100 | 105 | 112 | 119 | 126 | 134 | | | | | | |
| 75 | 84 | 88 | 92 | 97 | 103 | 109 | 116 | 124 | 132 | | | | | | | |
| 80 | 84 | 89 | 94 | 100 | 106 | 113 | 121 | 129 | | | | | | | | |
| 85 | 85 | 90 | 96 | 102 | 110 | 117 | 126 | 135 | | | | | | | | |
| 90 | 86 | 91 | 98 | 105 | 113 | 122 | 131 | | | | | | | | | |
| 95 | 86 | 93 | 100 | 108 | 117 | 127 | | | | | | | | | | |
| 100 | 87 | 95 | 103 | 112 | 121 | 132 | | | | | | | | | | |

# Heat Index Risks

Meteorologists use the Heat Index to warn the public of health risks in the summer. Interestingly, the Heat Index values in Table 4.3.1 are not in direct sunlight, but in the shade. Exposure to full sunlight can increase the heat index by up to 14 °F. Table 4.3.2 lists the effects of different heat index values on our health.

*Table 4.3.2: Effects of the heat index on the human body.*

| Heat index | Effects |
|---|---|
| 81–90 °F | Caution: fatigue is possible with prolonged exposure and activity. Continuing activity could result in heat cramps. |

| Heat index | Effects |
|---|---|
| 90–106 °F | Extreme caution: heat cramps and heat exhaustion are possible. Continuing activity could result in heat stroke. |
| 106–129 °F | Danger: heat cramps and heat exhaustion are likely; heat stroke is probable with continued activity. |
| over 129 °F | Extreme danger: heat stroke is imminent. |

It is suggested that during high Heat Index days, you should drink plenty of water and reduce physical activity, especially outdoors. Other precautions include staying in the shade, using air conditioning when possible, and wearing lightweight, light-colored clothing.

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# 4.4: Dew and Frost

## Dew

Clear, calm nights allow the maximum release of longwave radiation from Earth's surface and surrounding objects into space. If the surface temperature falls to the dew point, water vapor in the air can begin to condense on the surface. We notice this early in the morning (when the temperatures are the lowest), as liquid droplets form on cars, and on blades of grass, as depicted in Figure 4.4.1. On the other hand, cloudy skies reflect and absorb sunlight, re-emitting longwave radiation to the surface, thereby reducing the amount of cooling that would otherwise occur. Windy conditions enable surface air to mix with drier air aloft. Thus, heavier dew tends to occur when the wind is light rather than when it is strong. Next time you walk across the grass and find your shoes completely wet, take note of the conditions that allowed dew to form.



Figure 4.4.1: As the ground cools by radiational cooling, water vapor condenses to form dew in the form of tiny droplets of water on blades of grass. (CC BY-SA 3.0; Kesoff via Wikimedia Commons)

During winter, as nights are longer than days, early-morning ground temperatures can fall below freezing. When temperatures drop below freezing and reach the dew point or frost point, the ice on the ground is referred to as frost or frozen dew. "Frost" can form in two ways: Either by deposition or freezing. Depositional frost or white frost occurs when the dew point (now called the frost point) is below freezing. When this frost forms, the water vapor goes directly to the solid state. Depositional frost covers vegetation, cars, and other objects with ice-crystal patterns (treelike branching). If the depositional frost is thick enough, it resembles a light snowfall as depicted in Figure 4.4.2.

Frost that forms when liquid water freezes is best referred to as *frozen dew*. Initially, both the dew point and temperature are above freezing when dew forms. Longwave radiational cooling gradually lowers the temperature to at or below freezing during the night. Cold-air advection at night can also cause temperatures to drop below freezing. Once the temperature falls to freezing, the condensed dew droplets freeze. Frozen dew looks different from white frost. Frozen dew does not have the crystal patterns of white frost. White frost appears whiter, while frozen dew tends to look slicker and less distinct.



Figure 4.4.2: When the ground cools to temperatures below freezing, ice appears on the ground instead of water droplets and is referred to as frost or frozen dew. (Public Domain; Rosendahl via Wikimedia Commons)

Compared to dew, frost and frozen dew can cause morning transport delays by sticking to car windshields. Some frost or frozen dew is much easier to scrape off a car than others. When the temperature is near freezing (29 to 32°F), ice is relatively easy to scrape from car windows. It is also quicker to warm up the car windows to above freezing with the defroster when temperatures are near freezing. Once temperatures drop below 28°F, the ice becomes more difficult to remove. It requires more "elbow grease" to remove the ice. It also takes longer to warm up the car windows to above freezing. Next time you witness ice in the morning, think about the processes that produced the frost or frozen dew.

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# 4.5: Fog

## Radiation Fog



Figure 4.5.1: Early morning radiation fog in Bangladesh. (CC BY-SA 4.0; Abdul Momin via Wikimedia Commons)

Fog is often described as a cloud resting near the ground. Fog forms when the air's temperature and dew point are equal. This can occur either by cooling the air or by adding sufficient moisture to raise the dew point. When composed of ice crystals, it is called ice fog.

Radiation fog is a very common type of fog throughout the United States and worldwide. Figure 4.5.1 shows radiation fog forming in the early morning hours in Bangladesh. It is most prevalent during the fall and winter, forming overnight as the air near the ground cools and stabilizes. When this cooling causes the air to reach saturation, fog will form. Fog will first form at or near the surface, thickening as the air continues to cool. The fog layer will deepen overnight as the air above it cools. As this air cools, the fog will extend upward. The most favored areas for fog development are sheltered valleys with little to no wind and locations near bodies of water. Wind would disrupt the formation of radiation fog. Radiation fog is usually patchy, tends to stay in one place, and goes away the next day under the sun’s rays. Thicker instances of radiation fog tend to form in valleys or over calm bodies of water.

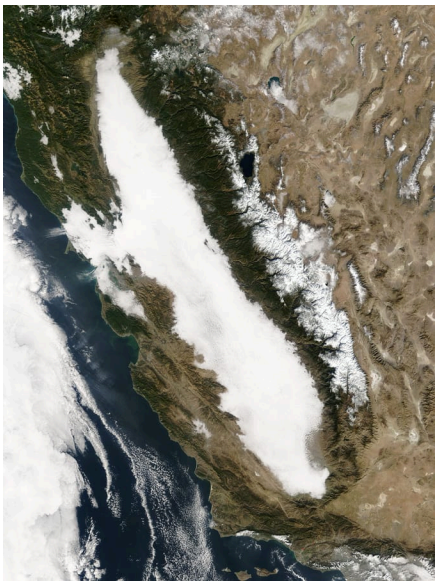


Figure 4.5.2: Tule fog captured by NASA's Terra Satellite in January 2005. The Sacramento Valley and San Joaquin Valley are almost completely covered in tule fog. (Public Domain; Jeff Schmaltz via NASA)

Valley fog is a type of radiation fog. When air along ridgetops and the upper slopes of mountains cools after sunset, it becomes denser and heavier, draining into the valley floors below. As the air on the valley floor continues to cool through radiational cooling, it becomes saturated, and fog forms. Valley fog can be very dense at times. It is also called “tule” (TOO-lee) fog and occurs each winter in the Central Valley of California. This type of fog tends to dissipate quickly once the sun rises and begins to evaporate the fog layer. But under favorable conditions, it can persist for days or even weeks, as during the 2025 Tule Fog event in the California Central Valley. The combination of a cool, moist layer of air from the Pacific near the surface, clear skies above, and light winds results in exceptionally thick fog on many nights from late October to February. This phenomenon is often visible from satellite images like Figure 4.5.2.

## Advection Fog



Figure 4.5.4: Advection fog moving across the Golden Gate Bridge in San Francisco. (CC BY 2.0; Runner310 via Flickr)

Advection fog often resembles radiation fog and results from condensation. However, condensation in this case is caused not by a reduction in surface temperature but by the horizontal movement of warm, moist air over a cold surface, such as warm, humid air flowing over snow. Advection fog can sometimes be distinguished from radiation fog by its horizontal motion along the ground. It is most common at sea when moist air encounters cooler waters, including areas of cold water upwelling, such as along the

California coast. A sufficiently strong temperature difference between the Pacific Ocean and the coastal waters or bare ground results in persistent morning advection fog near San Francisco, as observed in Figure 4.5.4.

## Evaporation/Mixing Fog

This type of fog forms when sufficient water vapor is added to the air by evaporation and the moist air mixes with cooler, relatively drier air. A common example is steam fog, which forms when cold air moves over warm water, as depicted in Figure 4.5.5. When cool air mixes with warm, moist air over water, the moist air cools until its humidity reaches 100%, and fog forms. This type of fog appears as wisps of smoke rising from the water's surface. This type of fog commonly occurs over the Great Lakes but can also occur on any lake or body of water, as well as on wet surfaces. Steam fog does not penetrate very deeply, but it is sufficient to block some sunlight.



Figure 4.5.5: When warm, moist air rising over a lake comes in contact with the cold air, steam fog forms. (CC BY 2.0; Michael Scialdone via Flickr)

Another common example is breath fog. When we exhale, our breath is warm and contains a lot of water vapor. On a cold day, the cold air rapidly lowers the temperature of our breath, causing it to reach the dew point. It creates a misty cloud, which we have experienced during cold weather.

## Upslope Fog

This type occurs when sloping terrain lifts air, cooling it adiabatically to its dew point and saturation. Upslope fog may be viewed as either a stratus cloud or fog, depending on the point of reference of the observer. Upslope fog generally forms at higher elevations and moves downslope into valleys. This fog can persist at higher wind speeds due to increased lift and adiabatic cooling. Upslope winds of more than 10 to 12 knots usually result in stratus rather than fog. The east slope of the Rocky Mountains is a prime location for this type of fog.

## Freezing Fog




Figure 4.5.6: When the temperature falls to 32°F (0°C) or below, the tiny droplets in the fog freeze when they come into contact with an object. (Public Domain; Freezing Fog via NWS)

Tiny, supercooled liquid water droplets in fog can freeze instantly on exposed surfaces when surface temperatures are at or below freezing. Some surfaces that these droplets may freeze on include stairs and rails, sidewalks, roads, vehicles, and tree branches, as depicted in Figure 4.5.6. Extreme caution should be taken if travel is necessary. Freezing fog can cause black ice to form on roadways. Black ice is difficult to see, making it particularly dangerous. Drive more slowly when you suspect icy conditions. For those flying, a thin layer of ice can form on aircraft, making flights very dangerous unless the aircraft is treated or equipped with effective de-icing equipment.

## Ice Fog



Figure 4.5.7: A tree in a field during extreme cold with frozen fog. (CC BY-SA 3.0; Ian Furst via Wikimedia Commons)

Frozen fog (also known as ice fog) is any fog in which droplets have frozen into extremely tiny crystals of ice in midair. Generally, this requires temperatures at or below –35°C (–31°F), making it common only in and near the Arctic and Antarctic regions, as depicted in Figure 4.5.7. It is sometimes observed in urban areas, where it forms from the freezing of water vapor present in automobile exhaust and in combustion products from heating and power generation. Urban ice fog can become extremely dense and will persist day and night until the temperature rises. It can be associated with the diamond dust form of precipitation, in

which very small ice crystals form and slowly fall to the ground. This often occurs under blue-sky conditions, which can produce a variety of halos and other effects of sunlight refraction by airborne crystals, such as light pillars.

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# 4.6: Clouds

## Cloud types

Clouds form when the air reaches saturation high in the atmosphere. We learned earlier that in the Troposphere, temperature decreases with altitude. Thus, if warm, moist air near the surface were to rise adiabatically, it would cool as it expanded and reach its dew point, forming a cloud. Depending on the temperature within the cloud, they're composed of liquid droplets or ice crystals. Figure 4.6.1 provides a brief overview of the various cloud types and their common abbreviations.

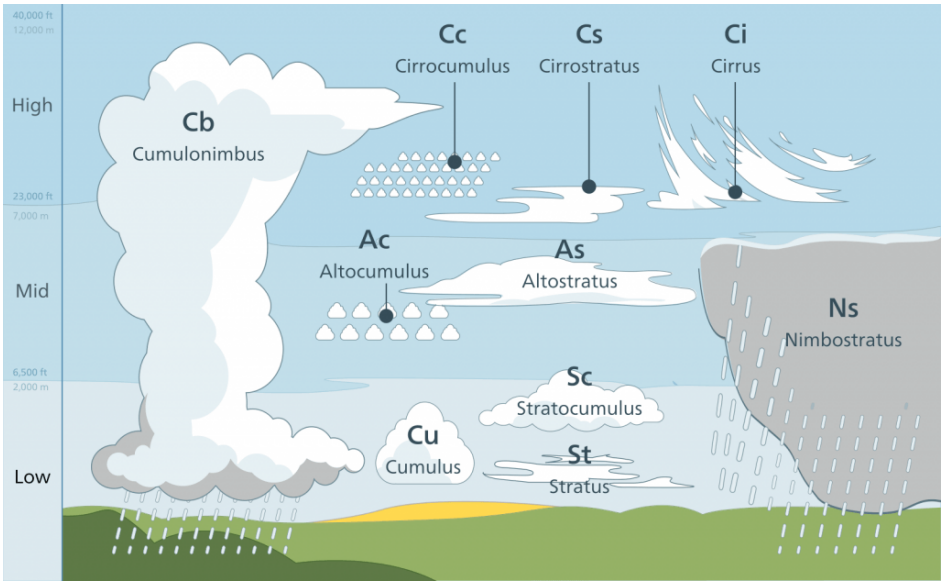



Figure 4.6.1: The cloud naming convention divides the atmosphere into the following cloud categories based on altitude: Low Clouds (below 6,500 ft) include Stratus (St), Stratocumulus (Sc), and Cumulus (Cu); Middle Clouds (6,500 to 20,000 ft); Altocumulus (Ac), Altostratus (As), Nimbostratus (Ns); and High Clouds (above 20,000 ft): Cirrus (Ci), Cirrocumulus (Cc), Cirrostratus (Cs). Nimbostratus (Ns) and Cumulonimbus (Cb) are rain-producing clouds that may cross the altitude boundaries between the classification levels. (CC BY-SA 3.0; Valentin de Bruyn via Wikimedia Commons).

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

## By Appearance and Shape

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

**Stratiform** clouds are horizontally layered clouds. They tend to spread over wide areas and take on a sheet-like appearance. 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 composed entirely of ice crystals are called “cold clouds” and typically have fuzzy edges. The edges appear fuzzy and not

well-defined because it takes longer for the ice phase to transition to the vapor phase than in warm clouds. This timescale results in a larger cloud-to-dry air transition region. High-altitude clouds are always cold.

Clouds composed of liquid water and ice crystals are called “mixed phase clouds”. It isn't easy to tell by eye whether a cloud is mixed-phase. Often, the decision comes down to the height of the cloud, since clouds that extend from the surface to high in the atmosphere likely contain a mixture of liquid droplets and ice crystals.

## By Altitude

Clouds at a high altitude (above 7 km) have the prefix “cirro” or “cirrus”. Cirrus clouds are made of ice crystals and often appear in the form of white, delicate filaments, mostly in patches or narrow bands. They may have a fibrous (hair-like) and/or silky sheen appearance as shown in Figure 4.6.2.



Figure 4.6.2: Cirrus Clouds over Indonesia (CC BY-SA 4.0; Stefy1576 via Wikimedia Commons).

Thin and white, cirrocumulus clouds look like a patchy sheet or layer arranged somewhat regularly into grains or ripples without shading, as shown in Figure 4.6.3.



Figure 4.6.3: Cirrocumulus clouds on a summer afternoon (CC BY-SA 3.0; King of Hearts via Wikimedia Commons).

Cirrostratus clouds appear as transparent, whitish, milky, veil-like clouds with a fibrous (hair-like) or smooth appearance, as shown in Figure 4.6.4. A sheet of cirrostratus is very extensive and can cover the whole sky, while still allowing sunlight through and creating shadows. The sun appears with halos around it because light reflects and refracts off ice crystals.



Figure 4.6.4: Cirrostratus clouds covering the clouds in a veil. (Public Domain; Ten Basic Clouds via NOAA).

Clouds at mid-altitude, between 2 and 7 km, have the prefix “alto”. They are usually made out of liquid droplets, but can be a mixture of liquid droplets and ice crystals. Altocumulus clouds appear as puffy masses, larger than cirrocumulus, as shown in Figure 4.6.5. They are also darker and more shaded than cirrocumulus clouds.

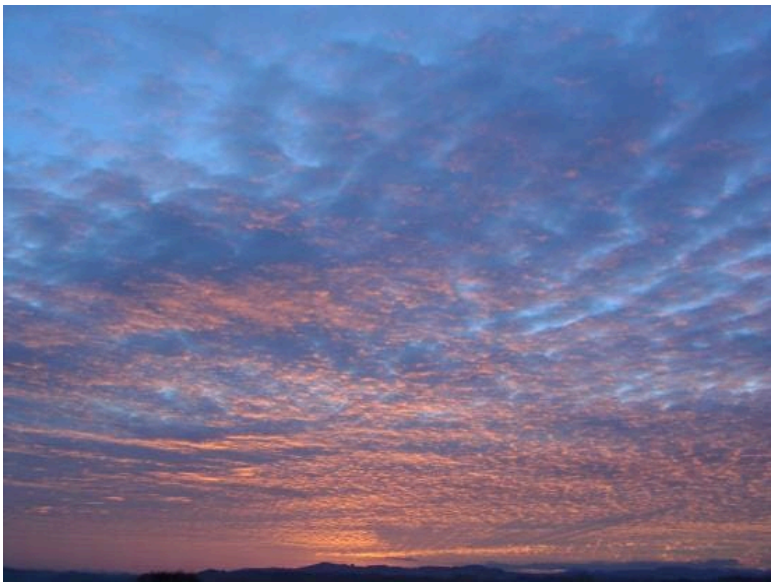


Figure 4.6.5: Altocumulus clouds often create beautiful sunset skies. (CC BY-SA 3.0; Simon Eugster via Wikimedia Commons)

Mid-level clouds can also form layers and are called Altostratus clouds, as shown in Figure 4.6.6. Altostratus clouds do not produce a halo phenomenon, nor are the shadows of objects on the ground visible. The sky is visibly darker, with a faint sun visible through the clouds.



Figure 4.6.6: Sun shines dimly through altostratus cloud layer. (CC BY-SA 4.0; Famartin via Wikimedia Commons)

Low-level clouds generally have the prefix "strato", which means layered. Stratocumulus clouds are gray, patchy, sheet, or layered clouds that almost always have a dark honeycomb appearance with rounded masses or rolls. The individual masses are much larger than altostratus or cirrostratus and much darker, as shown in Figure 4.6.7.



Figure 4.6.7: Stratocumulus clouds appear much darker, with regularly arranged masses. (Public Domain; Ten Basic Clouds via NOAA)

## By Characteristics

The prefix “nimbo” or suffix “nimbus” indicates a precipitating cloud. Nimbostratus usually has light to moderate precipitation as shown in Figure 4.6.8. It is thick enough throughout to blot out the sun. Low, ragged clouds called *stratus fractus* frequently occur beneath this cloud and sometimes merge with its base.



Figure 4.6.8: Nimbostratus clouds cover the whole sky and produce rain. (CC BY-SA 3.0; Togabi via Wikimedia Commons)

A cumulonimbus or thunderstorm cloud is a heavy and dense cloud in the form of a mountain or huge tower. The upper portion is usually smoothed and flattened in the shape of an anvil or vast plume due to strong upper-level winds, as shown in Figure 4.6.9. Some of the strongest updrafts can be found in the core of this cloud. Cumulonimbus clouds produce rain, thunder, lightning, and can sometimes also produce hail and tornadoes.



Figure 4.6.9: A cumulonimbus cloud rising tall with a flattened top. (Public Domain; NOAA via Flickr)

## Other

The above clouds represent the primary cloud types; however, many other cloud names and types are used in atmospheric sciences. A few common cloud names are listed below.

*Lenticular clouds* are stationary lens-shaped clouds with a smooth appearance that usually form 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. As air rises over a mountain, it cools and forms a cloud. As it descends on the other side, it heats up, and the cloud dissipates, leaving the lenticular cloud near the summit, as shown in Figure 4.6.10.



Figure 4.6.10: A lenticular cloud covers the summit crater of Mayon Volcano, Philippines. (CC BY-SA 4.0; Patryk Reba via Wikimedia Commons)

*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, as shown in Figure 4.6.11.



Figure 4.6.11: Mammatus clouds over the Nepal Himalayas. (CC BY-SA 4.0; Anton Yankovyi via Wikimedia Commons)

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# 4.7: Exercises

## Chapter 4: Exercises

1. Explain the conditions needed for saturation to occur.
2. Explain the difference between absolute humidity, specific humidity, and relative humidity.
3. If the ambient temperature is 90°F and the relative humidity is 75%, what is the heat index?
4. What are the health risks posed by high Heat Indices (over 80)?
5. Define Wet bulb temperature. How is it measured?
6. Distinguish between Dew and Frost.
7. Name 4-5 different types of fog.
8. Name different cloud types by appearance, phase, and altitude.

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# 5: ATMOSPHERIC STABILITY AND PRECIPITATION

## Introduction

Have you ever experienced a time where weather conditions quickly deteriorated, going from sunny to cloudy, to stormy in a short period of time? While there are many potential causes of sudden, drastic weather changes, one of the most common is atmospheric instability. This condition allows air to rise rapidly, then cool and condense into rainfall. On the other hand, when the atmosphere is stable, air cannot rise, allowing air pollutants from fires, factories, and human sources to settle near the surface, resulting in poor air quality. In either case, atmospheric stability is an important factor in local weather conditions and is of concern to weather forecasters and researchers. In this chapter, we will cover the fundamentals of atmospheric stability, how to determine atmospheric stability, and then discuss precipitation formation processes, which occur when rising air cools and condenses.



Figure 5.1: Cumulonimbus clouds can form in unstable weather conditions. (CC-BY-SA 4.0; Jakob Halun via Wikimedia Commons)

### 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
- Distinguish warm and cold clouds and the precipitation formations that occur in both types of clouds.


5.1: What is Stability?

5.2: Atmospheric Stability and Lapse Rates

5.3: Cloud Droplets

5.4: Precipitation Types

5.5: Exercises for Chapter 5



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# 5.1: What is Stability?

## Stable vs. Unstable



Figure 5.1.1: A cumulonimbus cloud rises and expands as a product of instability within the atmosphere (Public Domain; NASA via Wikimedia Commons).

When you think of the word “stable,” you typically think of an object that is unlikely to change or something 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 it changing shape from one minute to the next. Such clouds are constantly changing and thus represent the atmosphere in an unstable state.



Figure 5.1.2: A perfect cumulus cloud just west of Magic Island (CC-BY-SA 4.0; Sarah Williamson via Atmospheric Processes and Phenomenon).

Atmospheric instability is a concept closely linked to thunderstorms, cumulus development, and vertical motion. To visualize the

concept of **stability**, imagine a boulder at the bottom of a canyon, surrounded by steep hills, as depicted in Figure 5.1.3 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. In a stable atmosphere, the temperature of a parcel is cooler than that of the atmosphere, so any air (or clouds as depicted in Figure 5.1.3) would be forced to sink. To visualize the concept of **instability**, imagine the same boulder at the top of a hill (red circle). 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. In an unstable atmosphere, the temperature of an air parcel is warmer than that of the surrounding atmosphere, and the parcel (and any associated clouds, like depcited in Figure 5.1.3) would be forced to rise. However, if the same boulder were to be placed on flat ground (green circle) and you were to push it, it would change position, but remain in its new position. This is an example of **neutral stability**. In a neutral atmosphere, a parcel of air has the same temperature as the atmosphere, and any air (or clouds, as depicted in Figure 5.1.3) would neither rise nor sink.

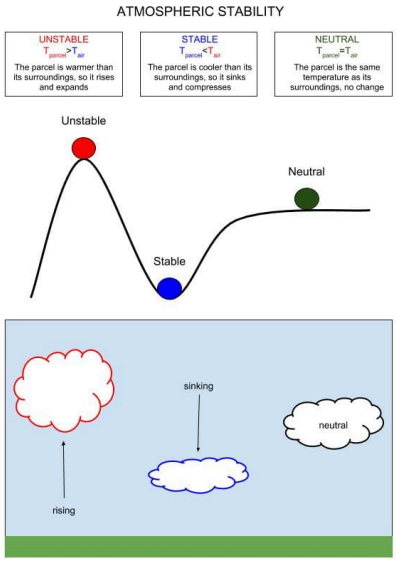



Figure 5.1.3: Examples of stability and instability in relation to air and parcel temperatures (CC-BY-SA 4.0; Britt Seifert via Atmospheric Processes and Phenomenon).

Each of these concepts can be applied to motions of air parcels in the atmosphere. Stability in atmospheric science is important because cloud formation is closely linked to atmospheric stability or instability. In this chapter, we will connect these concepts to the buoyancy of air parcels and learn to use thermodynamic diagrams to visualize movement.

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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 point is that the movement of air parcels in the atmosphere can be approximated 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 region where the surrounding air pressure is lower, because pressure decreases with height. Because the air pressure outside the parcel is lower than the pressure inside, the air molecules inside the parcel will effectively push outward on the parcel's walls, causing it to expand adiabatically. The air molecules inside the parcel must use some of their own energy to expand the air parcel's walls. Hence, 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. Higher environmental pressure will push inward on the parcel walls, compressing them and raising 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 they are not due to mixing; they are due to changes in the air parcel's internal energy.

## 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;** when it decreases with altitude, it is defined as positive. **The dry adiabatic lapse rate is equal to 10°C/km**. This drop in temperature is due to adiabatic expansion and a decrease in internal energy. An example of how dry air cools as it rises is provided in Figure 5.2.1:

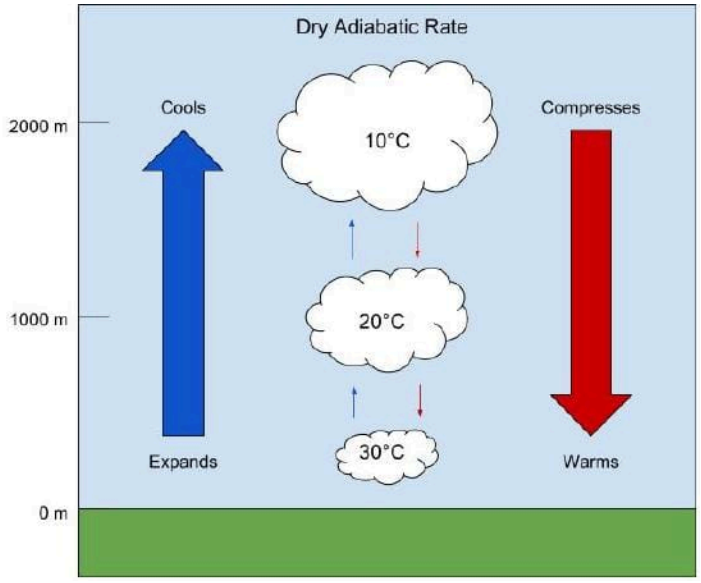



Figure 5.2.1: Air rises, expands, and cools at the dry adiabatic lapse rate, approximated as a 10°C decrease per km (CC-BY-SA 4.0; Britt Seifert via Atmospheric Processes and Phenomenon).

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 a boulder on a hill or in a valley, some initial movement resulted in either more (unstable), less (stable), or no change (neutral). Given an initial change in the elevation of an air parcel, if the air is in stable equilibrium, the parcel will tend to return to its original position after being 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 favor 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 whether an air parcel will remain stable after an 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 10°C/km, 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 a new equilibrium where its temperature matches the environment's. In this case, because an initial change is amplified, the air parcel is unstable. To determine whether the air parcel is unstable, we need to know the temperatures 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 transmits data on temperature, humidity, wind speed, and position, which are plotted on a thermodynamic diagram. This vertical plot of temperature and other variables is known as a **sounding**.

## 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 Clausius-Clapeyron graph shows that, for a given amount of moisture, air is more likely to be saturated at lower temperatures.

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 into liquid water, forming a cloud.

When water vapor condenses, it transitions 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 is released as latent heat. The latent heat of condensation is approximately equal to $2.5 * 10^6$ $J \cdot kg^{-1}$, 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 significant 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 during 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 slightly with temperature, but we will treat it as a constant for simplicity: **6°C/km.**

## Determining if a Parcel is Saturated and Dew Point Lapse Rates

A common question when discussing the dry and moist adiabatic lapse rates is "Which one should I use?" The answer to that question is primarily determined by whether the parcel is saturated. While there are many different technical ways to determine the saturation status of a parcel, the simplest way (and the one we'll use here) is to compare the temperature of our parcel to the dew point temperature:

- A Parcel is **unsaturated** (hence, you would use the Dry Adiabatic Lapse Rate) if the temperature of the parcel is **warmer than the dew point**.
- A Parcel is **saturated** (hence, you would use the Moist Adiabatic Lapse Rate) if the temperature of the parcel is **equal to the dew point**.
- When a parcel's temperature is below its dew point, it is said to be "supersaturated." Super saturation is a temporary state because excess moisture in the parcel quickly condenses, lowering the dew point to the parcel's temperature. Because of that, we won't worry about supersaturated parcels here.

Because a rising parcel experiences a decrease in air pressure, the dew point temperature decreases with height, and therefore it has its own lapse rate. Just like the parcel's air temperature, the parcel's dew point lapse rate is determined by saturation:

- If a Parcel is Unsaturated, the dew point's lapse rate is 2**°C/km**
- If a Parcel is Saturated, the dew point's lapse rate is 6**°C/km.**

This is because, as a saturated parcel cools, moisture condenses out of the parcel, causing the dew point to drop at the same rate as the air temperature.

## Moist Stability

Changes in moisture alter the air parcel's lapse rate and, therefore, affect 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—we need to know whether the air parcel is dry or moist. Some definitions are included in Figure 5.2.2, which take into account both dry and wet adiabatic lapse rates.

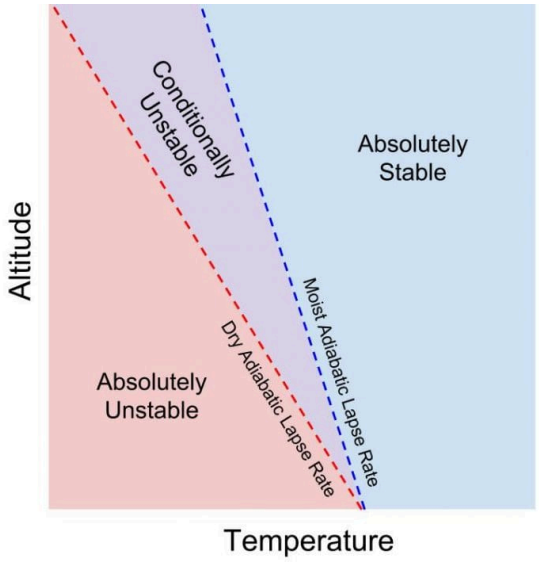



Figure 5.2.2: A thermodynamic diagram showing the stability of the atmosphere based on the dry (10 C $km^{-1}$) and moist (6 C $km^{-1}$) adiabatic lapse rates (CC-BY-SA 4.0; Britt Seifert via Atmospheric Processes and Phenomenon).

Rising air cools at two different lapse rates, depending on whether the parcel is staturated or unsaturated. An unsaturated parcel of air cools faster with height, thus the line for the **dry adiabatic lapse rate** is not as steep as the **moist adiabatic lapse rate**, which represents a saturated parcel cooling at a slower rate with height (hence a steeper line).

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 faster than its surroundings, even after it reaches saturation. If an air parcel is cooler at all levels, it cannot rise, even after it becomes saturated (when latent heating counteracts some cooling).

The atmosphere is said to be **absolutely unstable** if the environmental lapse rate exceeds 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 it will be warmer (and less dense) than the environment and will be 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 tendency of an air parcel to rise) depends on whether 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 becomes saturated, however, it will cool at the moist adiabatic lapse rate. In this case, it will cool more slowly than the environment, then warm up, and rise.

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# 5.3: Cloud Droplets

## Aerosols

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 atmospheric particles known as aerosols, or, more generally, cloud condensation nuclei (**CCN**). Aerosols are everywhere in the atmosphere. They can be composed of earthy matter, such as 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, such as sulfates and nitrates, volatile organic compounds, or even phytoplankton. Anything small enough to be lofted and suspended in the atmosphere can become an aerosol. If there were no aerosols in the atmosphere, relative humidities well over 100% would be needed for clouds to form. In laboratory studies under a clean atmosphere (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.

Many CCNs are hygroscopic or "water-loving", meaning they tend to absorb moisture, so condensation may start before the relative humidity reaches 100%. For example, when condensation occurs on extremely hygroscopic salt particles, it 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 lifted by a mountain or a rising thermal, it would cool, and the relative humidity would increase. As air nears saturation, condensation will begin on the largest and most hygroscopic CCNs. Soon, a cloud of many small cloud droplets, far too small to fall at any significant speed, will form.

## Raindrops

So how can cloud droplets grow to form raindrops? The condensation process is insufficient and 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 theories of rain formation.

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

## Collision-Coalescence Process

In warm clouds, where all cloud droplets are liquid, the **collision-coalescence** process is the primary mechanism for producing precipitation. This is thought to be especially true 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 ratio. This advantage allows them to fall faster and collide with smaller cloud droplets. Sometimes cloud droplets stick together and coalesce into a larger droplet. This begins a positive feedback loop in which these larger droplets fall even faster, collide with even more smaller droplets in their path, and aggregate more and more cloud droplets. However, note that a collision between cloud droplets does not always mean that coalescence will occur. Sometimes droplets will bounce apart during collisions if their surface tension is too high. For collision-coalescence to begin, a cloud needs to have a wide distribution of cloud droplet sizes. This can occur from variations 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 a cloud droplet spends in a cloud, influences how large it can grow through collision-coalescence. The cloud height is, of course, a factor here, but it's a little more complicated than that. Rising motion in a cloud will slow the downward speed of a falling droplet. This can increase the time a cloud droplet spends in a cloud and the size it grows to. Let's cover a few examples.

Deep cumulus clouds with strong convective updrafts tend to produce larger raindrops because the updrafts are strong and droplets have more 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 able to grow very large. If moist air lies 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 reaching the ground.

To summarize, in warm clouds, cloud droplets grow to precipitation-sized drops through collision-coalescence. 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. Figure 5.3.1 illustrates the collision-coalescence process of raindrop production.

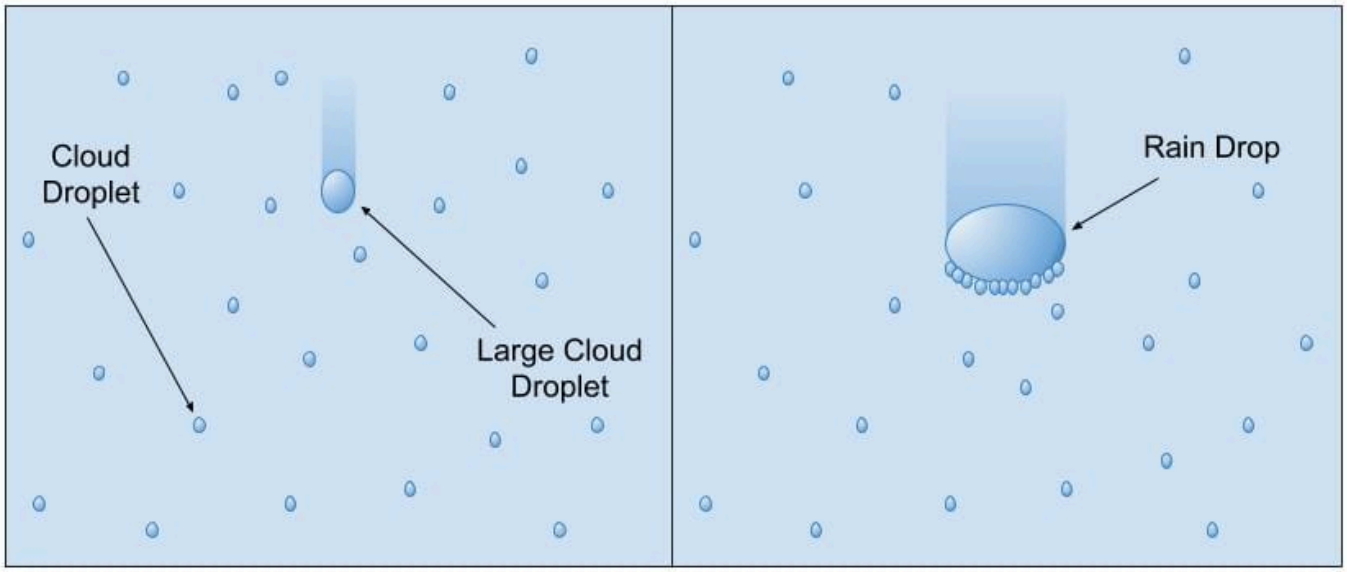



Figure 5.3.1: The collision-coalescence process occurring in warm clouds. The left image shows a slightly larger spherical cloud droplet falling in an environment of smaller cloud droplets and demonstrates the importance of a range of cloud droplet sizes to initiate the collision-coalescence process. The image at right shows a fully formed oval-shaped raindrop rapidly collecting additional droplets as it falls, and highlights the acceleration of the collision-coalescence process once a raindrop forms (CC-BY-SA 4.0; Britt Seifert via Atmospheric Processes and Phenomenon).

## Ice Phase (Bergeron) Process



Figure 5.3.2: Ice crystals forming on a window at 30,000 feet (CC-BY-SA 3.0; Brocken Inaglory via Wikimedia Commons).

Outside 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, where temperatures are below 0°C. To understand why, we need to know how liquid water droplets freeze.

## Supercooled Water and Ice Nuclei

**Supercooled water** is liquid water below its freezing point (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 is similar to Figure 5.3.3.

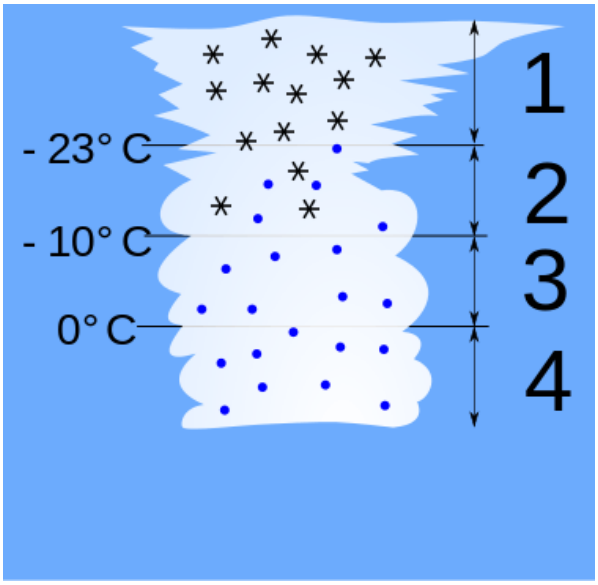



Figure 5.3.3: A simple diagram showing the distribution of liquid and solid phase hydrometeors in a mixed phase cloud. Description in the following text. (CC-BY-SA 3.0; BoH via Wikimedia Commons)

At low elevations above 0°C (region 4), the hydrometeors in the cloud exist as liquid droplets. Between 0°C and -10°C (region 3), supercooled liquid droplets exist. In the layer between -10°C and -23°C, some liquid droplets begin to freeze, and both liquid and ice-phase hydrometeors coexist (region 2). Finally, above some level where the temperature is below -40°C, all hydrometeors will exist in their solid state (region 1).

When liquid water droplets freeze without any nucleus, this is known as **homogeneous** 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, forming a tiny ice structure known as an *ice embryo*. When this embryo reaches a certain size, it will act as an ice nucleus. The other molecules in the droplet then become attached to the ice structure, and the entire droplet freezes.

Tiny ice embryos form when the in-cloud temperature drops just below freezing, but at these temperatures, thermal agitation is typically enough to weaken their structure and break them apart. At lower temperatures, there is less thermal motion, and ice embryos are more likely to grow 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 a 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. A common cloud-seeding agent is silver iodide (AgI).

Some IN allow water vapor to immediately become solid ice when they come into contact with each other. These are known as

*deposition nuclei* because water vapor changes phase directly into solid ice without first becoming liquid (the phase change from gas to solid is called deposition). INs that are effective at freezing supercooled liquid droplets are called **freezing nuclei**. Some freezing nuclei must be immersed in a liquid drop to freeze, while others are effective at inducing condensation and then freezing. Many freezing nuclei cause supercooled droplets to freeze upon collision, a process called **contact freezing;** these nuclei are called **contact nuclei**. These different freezing methods are outlined in Figure 5.3.4.

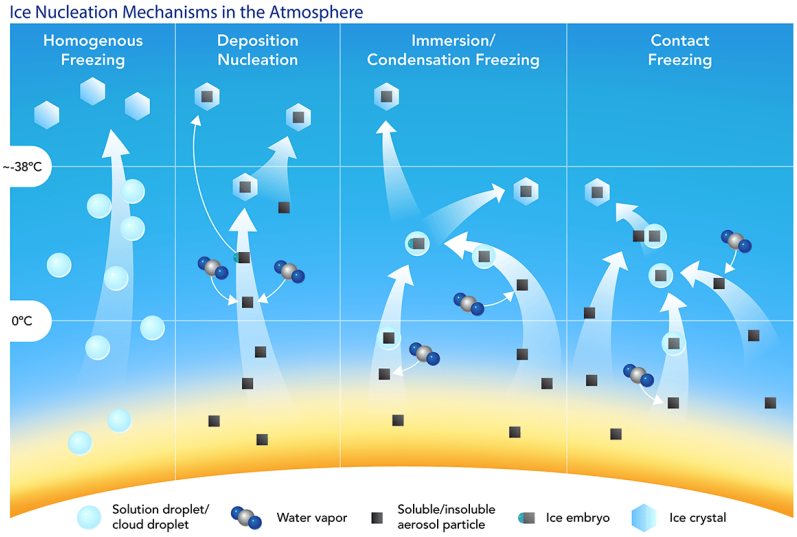



Figure 5.3.4: Four processes of ice formation in the atmosphere through four mechanisms: homogeneous freezing, deposition nucleation, immersion/condensation freezing, and contact freezing. Each process involves interactions among cloud droplets, water vapor, aerosol particles, and ice embryos, leading to the formation of ice crystals at temperatures below freezing. (CC-BY-SA 4.0; Pre2grk, via Wikimedia Commons).

A detailed description of Figure 5.3.4 is provided in Table 5.3.1.

*Table 5.3.1. Different mechanisms of frozen precipitation formation.*

| Mechanism | Temperature range | Initial phase | Trigger/Process | Resulting phase |
|---|---|---|---|---|
| Homogeneous freezing | ≤ ~−38°C | Liquid droplet | Spontaneous freezing without particles | Ice crystal |
| Deposition nucleation | < 0°C (often well below) | Water vapor | Vapor deposits directly onto aerosol particle | Ice crystal |
| Immersion freezing | < 0°C | Liquid droplet | Aerosol particle immersed in droplet induces freezing | Ice crystal |
| Condensation freezing | < 0°C | Water vapor → droplet | Condensation onto particle followed by freezing | Ice crystal |
| Contact freezing | < 0°C | Liquid droplet | Collision with aerosol particle triggers freezing | Ice crystal |

To summarize, cloud droplets may freeze spontaneously, but only at very low temperatures. Ice nuclei can help ice crystals grow, 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 supercooled 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 is 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 over liquid and ice differ slightly, the presence of this new ice crystal significantly impacts the cloud.

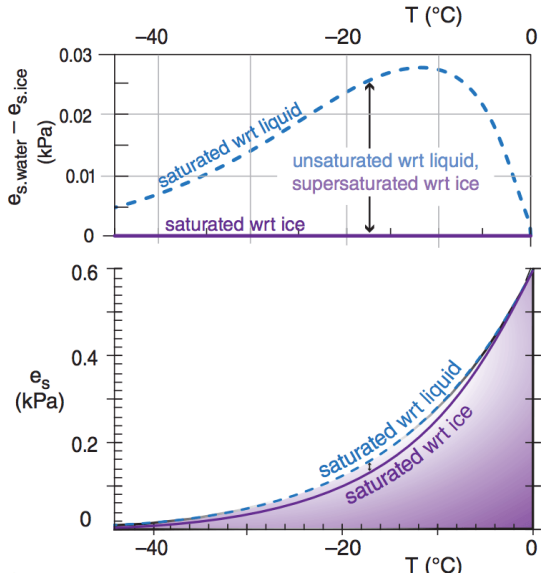



Figure 5.3.5: The saturation vapor pressure with respect to liquid (blue) and with respect to ice (purple). The difference is shown at the top (CC-BY-NC-SA 4.0; Roland Stull via Practical Meteorology).

Figure 5.3.5 compares water vapor saturation over liquid water and ice at different temperatures. The upper graph illustrates the difference in vapor pressure between water and ice, with regions labeled as saturated or unsaturated relative to each phase. The difference between saturation vapor pressure with respect to liquid and saturation vapor pressure with respect to ice starts out small (around 0.005 kPa) at a temperature below -40°C, but then gradually increases to a peak of aproximately 0.025kPa at a temperature of approximately -10°C before becoming 0 at 0°C. The lower graph depicts how saturation vapor pressure increases with temperature, with separate curves for saturation relative to liquid water and to ice. The difference between the two curves is calculated in the upper graph, but note that between approxmately -40°C and 0°C, saturation vapor pressure with respect to ice is lower than saturation vapor pressure with respect to liquid.

With respect to the liquid, the liquid droplets were at saturation. But now, with respect to ice, the ice crystal is in a supersaturated environment. 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, at a given temperature, many more molecules will escape from the liquid water surface, and it will require more water vapor around the droplet to keep it 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 between water and ice causes water vapor molecules to deposit onto the ice crystal from the environment. Because vapor molecules are removed from the surrounding environment, the vapor pressure above the water surface decreases. This throws the water droplets out of equilibrium with their surroundings, causing them to evaporate and replenish the water vapor removed 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 **Bergeron** (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 that 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 with updrafts and rising air apply, 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 the droplets to freeze onto the crystals. The ice crystal will grow larger and larger as it collides with 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 with other droplets, which in turn may also splinter, causing a chain reaction.

In colder clouds, ice crystals may collide 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**.

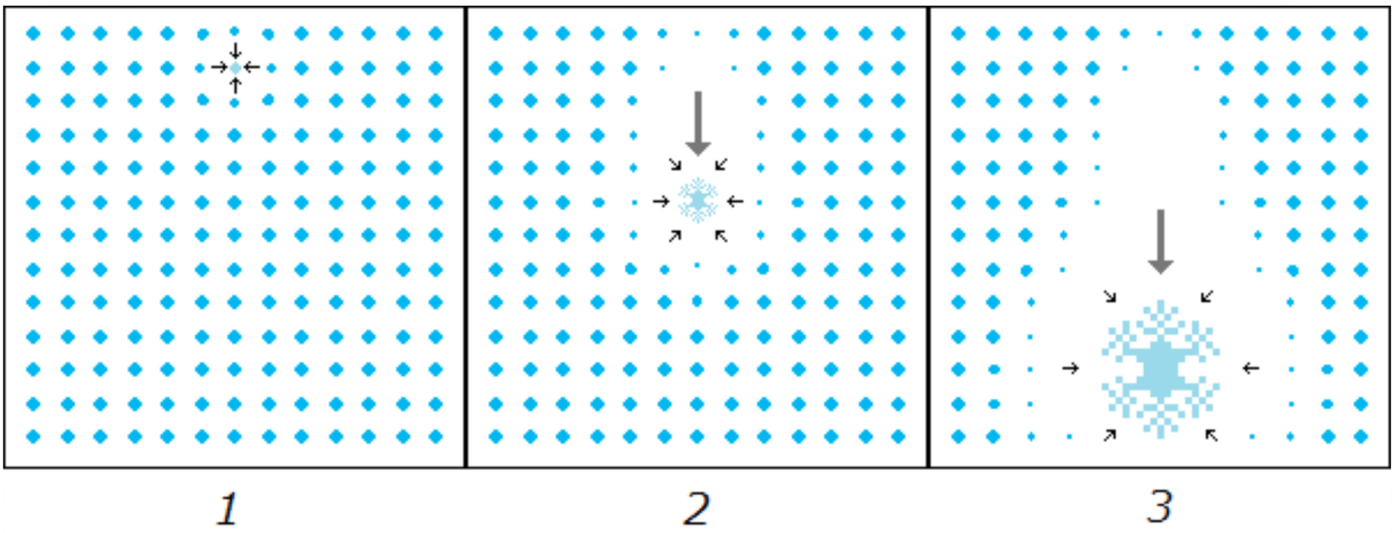



Figure 5.3.6: The growth of ice crystals occurs during their descent through aggregation (CC 0; Pekachu via Wikimedia Commons).

Figure 5.3.6 illustrates the formation and growth of an ice crystal in a supersaturated environment. In the first panel, a small disturbance occurs among water molecules. In the second, an ice embryo begins to form as molecules cluster. In the third stage, the embryo develops into a larger crystal as more water molecules join. 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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# 5.4: Precipitation Types

## Precipitation

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 falls through the atmosphere is called precipitation. However, we know that not everywhere in the world gets snow all the time. While the ice-phase process helps precipitate, 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 into a cold layer, usually just above the surface, and fall as frozen raindrops.

**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 a cloud.

As you can imagine, 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, we will learn how air masses and fronts combine regularly in Earth's atmosphere to create conditions that are conducive to all types of precipitation.

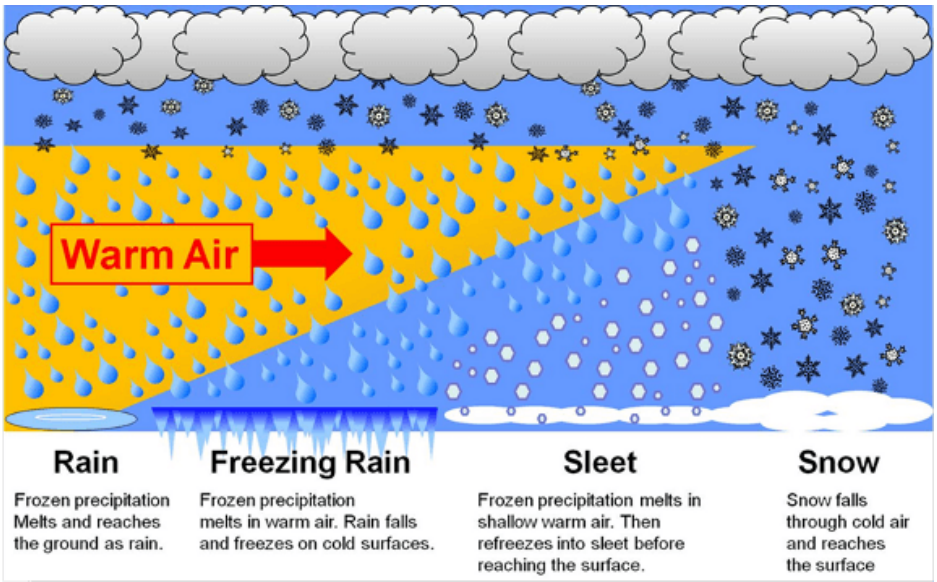



Figure 5.4.1: Precipitation changes associated with the passage of a warm front. Each precipitation type results from frozen precipitation interacting with layers of warm and cold air. Rain forms when frozen precipitation melts completely, freezing rain occurs when melted precipitation refreezes on cold surfaces, sleet forms when melted precipitation refreezes before reaching the ground, and snow occurs when precipitation remains frozen throughout its fall. (Public Domain; National Weather Service, via Wikimedia Commons).

Figure 5.4.1 provides an example of how a warm front creates temperature gradients in the atmosphere that produce precipitation types from rain, freezing rain, sleet, and snow, depending on your location with respect to the front.

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# 5.5: Exercises for Chapter 5

## Chapter 5: Exercises

1. Consider the following temperature-dew point combinations. Determine if the air is saturated or unsaturated.
    1. Temperature: 15°C Dew Point: 11°C
    2. Temperature: 11°C Dew Point: 11°C
    3. Temperature: 4°C Dew Point: -1°C
    4. Temperature: -3°C Dew Point: -3°C
2. Consider a parcel of air with a surface temperature of 11°C and a dew point of 3°C. Answer the following questions:
    1. Is the parcel saturated or unsaturated at the surface? Which lapse rates would you use for the parcel? Dew point?
    2. Raise the parcel to 1 km. What would its temperature and dew point be? Would the parcel be saturated? What lapse rates would you use to raise the parcel to 2 km?
    3. Raise the parcel to 2 km. What would its temperature and dew point be?
3. Suppose a weather balloon launched at the exact location and time as the parcel in question 2 showed an atmosphere with a temperature of 11°C and a lapse rate of 8°C/km. Answer the following questions:
    1. Is the atmosphere stable, conditionally unstable, or absolutely unstable? How do you know?
    2. Consider the parcel of air from question 2 at 2 km. Would it rise, sink, or stay in place? How do you know?
4. Which rain formation process produces most of the world's precipitation? What type of cloud does this process occur in?
5. What factors determine how large a cloud droplet can grow through collision-coalescence

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# 6: ATMOSPHERIC FORCES AND WIND

## Introduction

Wind is simply air in motion. In meteorology, when discussing wind, we are primarily concerned with its horizontal speed and direction. For example, if you hear a report of a west wind at 15 mph (24 km/h), that means the horizontal winds will be coming *from* the west at that speed. Although we cannot actually see the air moving, we can measure its motion by the force it exerts on objects. We use a wind vane to indicate the wind's direction and an anemometer to measure the wind's speed, as shown in Figure 6.1.



Figure 6.1: A weather station at a water treatment plant consisting of a wind vane and a cup anemometer. (CC BY-SA 3.0; William Demchick via Wikimedia Commons)

Newton's first law of motion states *that an object at rest will remain at rest unless acted upon by an external unbalanced force*. A Force is a pull or push that changes the resting state, motion, or direction of an object. Thus, as with all moving things, wind is caused by a net force acting on air. In this chapter, we'll examine the forces that act on air to cause winds.

### Learning Objectives

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

- Define wind and why it occurs
- Describe the four 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)

6.1: Wind

6.2: Force Balances

6.3: Exercises

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# 6.1: Wind

## Forces acting on air

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

1. Pressure gradient force
2. Coriolis force
3. Centrifugal force
4. Frictional force

### Pressure Gradient Force

Surface weather maps indicate regions of high- and low-pressure centers as H and L, as illustrated in Figure 6.1.1. Usually, surrounding these "highs" and "lows" are lines called isobars. "Iso" means "equal" and a "bar" is a unit of pressure, so an isobar means lines of "equal pressure". Along each line, the pressure is constant. In high-pressure systems, the air pressure along each isobar increases toward the center as the concentric lines are traced outward. The opposite is true for low-pressure systems, where each concentric line toward the center represents lower pressure. Isobars may be close together or far apart.

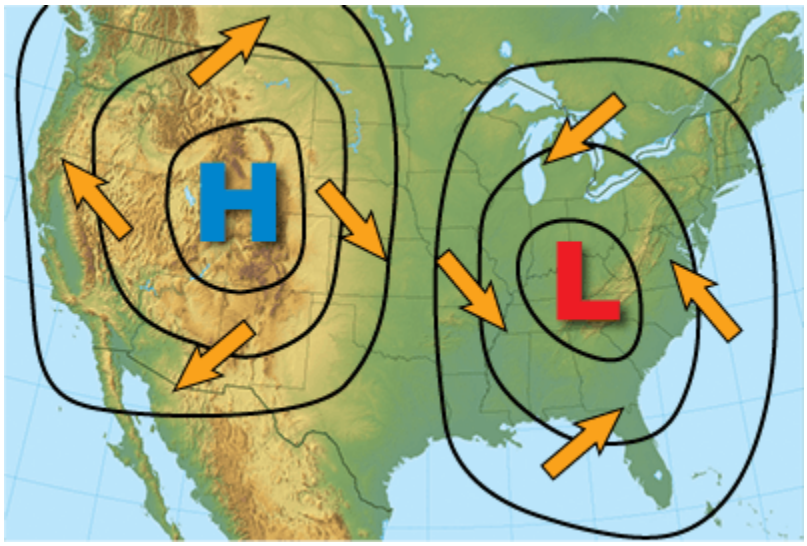



Figure 6.1.1: A depiction of a surface weather map with a high-pressure region over Utah, Colorado, and Wyoming, and a low-pressure region over Kentucky and Tennessee. Pressure contours are shown as lines of equal pressure, called isobars, and wind direction is clockwise and away from high pressure while counter-clockwise and into low pressure. (Public Domain; Origin of Wind via NOAA)

The P**ressure Gradient Force** ($PGF$) is the change in pressure over a distance. Therefore, the units of the pressure gradient force are Pascals per meter (P $m^{-1}$). The pressure gradient is the rate of change of pressure per unit distance. The magnitude of the pressure gradient determines the magnitude of the force it exerts.

$$PGF = \frac{\Delta P}{\Delta distance}$$

Without pressure differences, there would be no wind because there would be nothing to cause air to start moving. The speed of the wind is *directly proportional* to the pressure gradient, meaning that as the change in pressure increases (i.e., the pressure gradient increases), the speed of the wind also increases at that location.

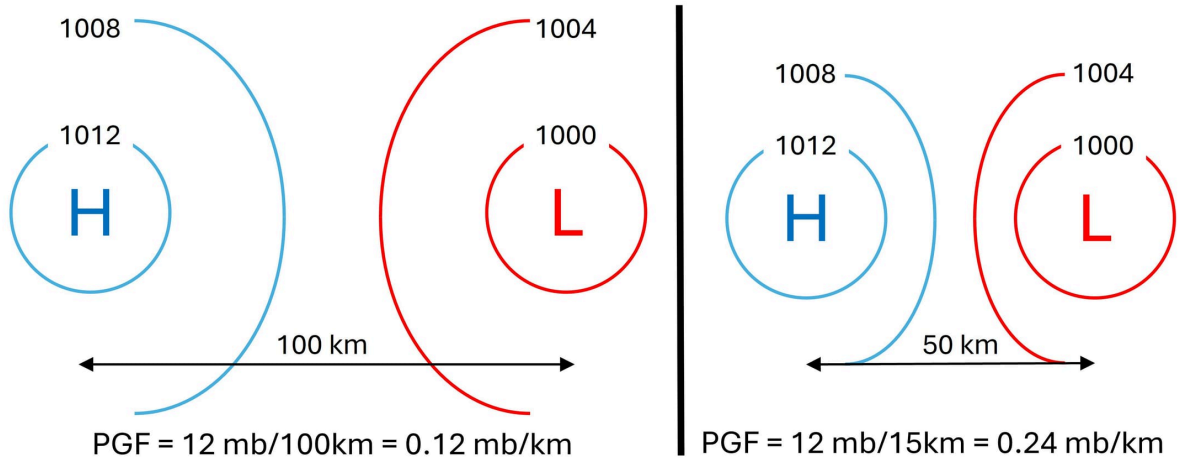



Figure 6.1.2: Isobars show a pair of high and low pressure regions separated by different distances. For the same pressure difference, the pressure-gradient force (PGF) is stronger when isobars are closer together. The high pressure is approximately 1012mb while the low pressure is approximately 1000mb. When the high- and low-pressure centers are separated by 100 km, the PGF is 12mb/100km = 0.12 mb/km. When the pressure centers are separated by 50 km, the PGF doubles to 12mb/50km = 0.24 mb/km. (CC BY 4.0; Neel Desai)

When the isobars are tightly packed, we know that there is a large change in pressure over a short distance as indicated in Figure 6.1.2. This suggests a strong pressure gradient force, resulting in higher wind speeds. When the isobars are farther apart, the pressure changes over a larger distance. This indicates a weaker pressure gradient force, resulting in lower wind speeds.

## Coriolis Force

The **Coriolis force** $(CF)$ is an apparent force that occurs due to the rotation of Earth. The Coriolis force is a deflecting force and acts only on objects already in motion. Therefore, it cannot generate wind, but it can deflect it and change its direction. 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, as illustrated in Figure 6.1.3.

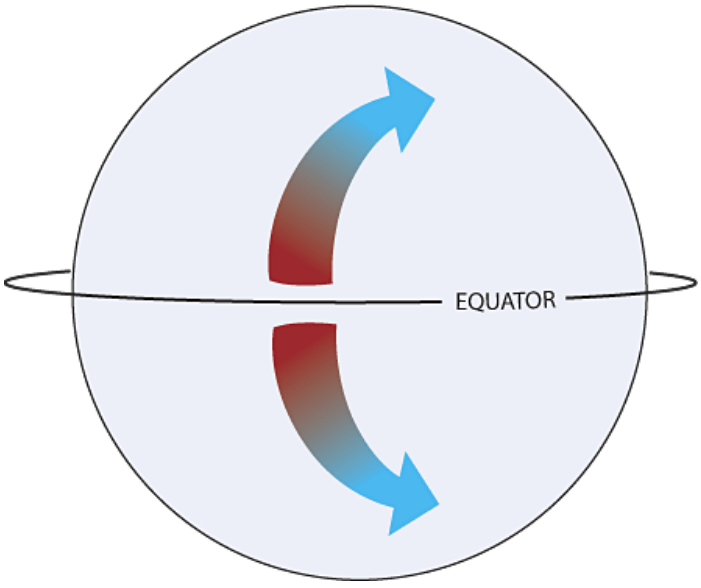



Figure 6.1.3: The rotation of the Earth on its axis deflects the atmosphere toward the right in the Northern Hemisphere and toward the left in the Southern Hemisphere, resulting in curved paths. (Public Domain; The Coriolis Effect via NOAA)

One way to observe this force in action is to examine what happens when a straight line is drawn on a turning disk, as illustrated in Figure 6.1.4. Picture the Earth as a turntable, spinning counterclockwise. A ruler is placed over the turntable, and a pencil will move in a straight line from the center to the edge. Since the Earth (the turntable) is rotating, the line created will appear curved and not be parallel to the ruler.

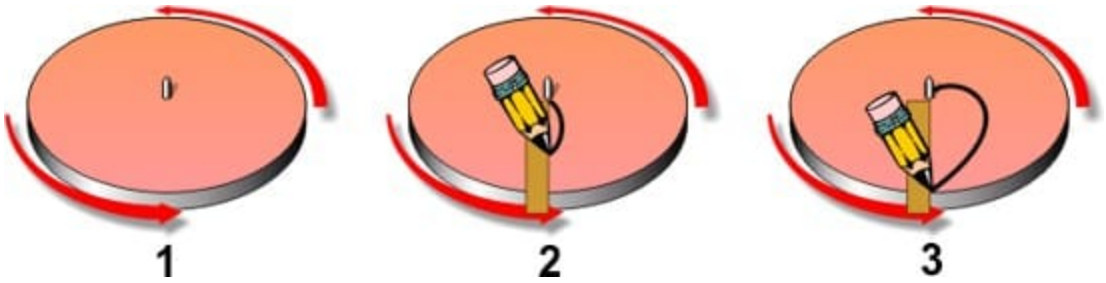



Figure 6.1.4: The Coriolis Force causes a straight-moving object to appear curved due to a rotating frame of reference. A straight line drawn using a pencil on a rotating disc appears curved. (Public Domain; Origin of Wind via NOAA)

Similarly, when viewed from space, wind appears to travel in a straight line. However, when viewed from Earth, air (as well as other objects in flight, such as planes and birds) is deflected to the right in the Northern Hemisphere and to the left in the Southern Hemisphere. The amount of deflection depends on:

- the rotation of the Earth
- the latitude
- the object's speed.

Thus, the deflection is greater at higher latitudes than at lower latitudes. The Coriolis force on a moving object on the equator is zero, while that at the poles is maximum. The Coriolis force is also 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.

## Centrifugal Force

The **Centrifugal force** is another apparent force that accounts for the effects of inertia when winds move along a curved path. The directionality of the centrifugal force points *outward* from the center of the curve. Centrifugal force is the opposite of centripetal force, which, in the case of winds, is often the Pressure Gradient Force. As we know, inertia is the tendency of an object to remain at rest or in motion. Therefore, inertia causes an air parcel to “want” to move along a straight line. Turning an air parcel along a curved path requires a centripetal force that acts toward 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.

## Frictional Force

Frictional force in the wind is a drag force, caused by the roughness of Earth's surface, that slows wind. The Frictional Force only acts on wind below 1000 m in altitude and always acts in the direction opposite to the direction of motion. If there is no movement, friction is zero. The magnitude of the Frictional force also depends on the type of surface. Imagine walking on a frozen lake. Friction is low on a frozen lake, allowing winds to flow unimpeded. However, winds over forest or mountainous terrain experience greater frictional drag. Near the surface, friction reduces the wind speed. At higher altitudes, farther from the surface, winds flow faster.

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# 6.2: Force Balances

## Geostrophic Balance

The four forces affect horizontal wind speed and direction, resulting in several common force balances throughout Earth's atmosphere. Geostrophic balance is arguably the most important force balance in the atmosphere and holds nearly all the time, except for a few specific cases, as discussed later. When the pressure-gradient force and the Coriolis force are equal and opposite, they balance each other out, and the wind is said to be in Geostrophic Balance. In **geostrophic balance**, $PGF = CF$. The resulting wind is called a geostrophic wind, and it flows parallel to isobars. Because geostrophic wind speed depends on the pressure gradient, winds are faster when isobars are closely spaced. In the Northern Hemisphere, the wind direction is parallel to the straight isobars, with low pressure to the left of the wind. In the Southern Hemisphere, the direction is parallel to the straight isobars with the low pressure to the wind's right. Figure $6.2.1$ shows the force balance present in a geostrophic wind in the northern hemisphere.

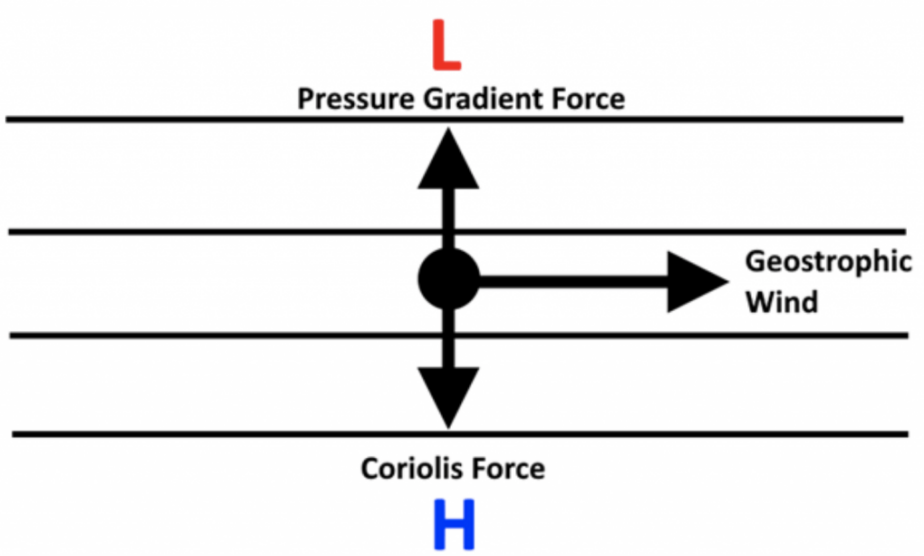



Figure $6.2.1$: Geostrophic wind force balance in the northern hemisphere includes the Pressure Gradient Force pulling the wind towards the low pressure and the Coriolis Force deflecting it to the right. As a result, geostrophic wind flows parallel to the isobars. (CC BY-SA 4.0; Shintaro Russell via Atmospheric Processes and Phenomena).

To achieve geostrophic balance, moving air undergoes **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 is in balance between the $PGF$ and the $CF$, moving parallel to the isobars rather than across them.

## Gradient Wind

The 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 when they are curved, but a different name is needed because the force balance includes an additional component. Compared to geostrophic winds, gradient winds feature a balance between the Coriolis force, the pressure gradient force, and the centrifugal force. Centrifugal force arises because the air is flowing along a curved path. The centrifugal force acts in the same direction as the Coriolis force, and the pressure gradient force acts in the opposite direction, as shown in Figure $6.2.2$.

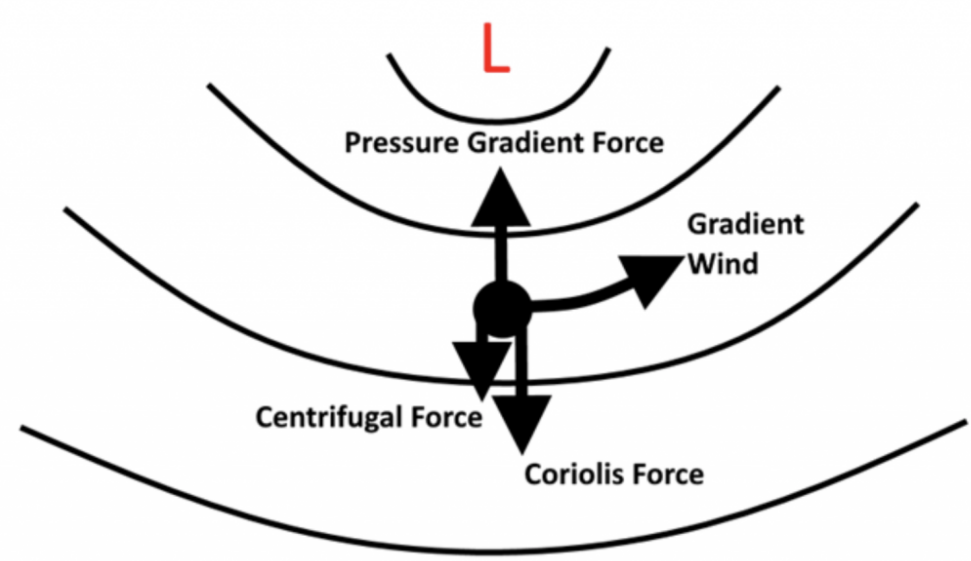



Figure 6.2.2: Gradient wind force balance includes the Pressure Gradient Force pulling the wind towards the low pressure, the Coriolis Force, and the Centrifugal force deflecting it to the right and opposite the low pressure. As a result, Gradient wind flows along curved isobars. (CC BY-SA 4.0; Shintaro Russell via Atmospheric Processes and Phenomena)

## Atmospheric Boundary Layer

Balanced wind near the surface occurs when the pressure-gradient force, the Coriolis force, and the frictional drag force are in balance. The frictional drag force acts in the opposite direction to the wind's motion, reducing wind speed at altitudes below 1000 m. The pressure gradient force doesn't change, but because the wind speed is slower, the Coriolis force is weaker. When that happens, the Coriolis force cannot balance the pressure-gradient force. Thus, the wind is pulled more by the pressure gradient and turns toward the low-pressure system, as shown in Figure 6.2.3.

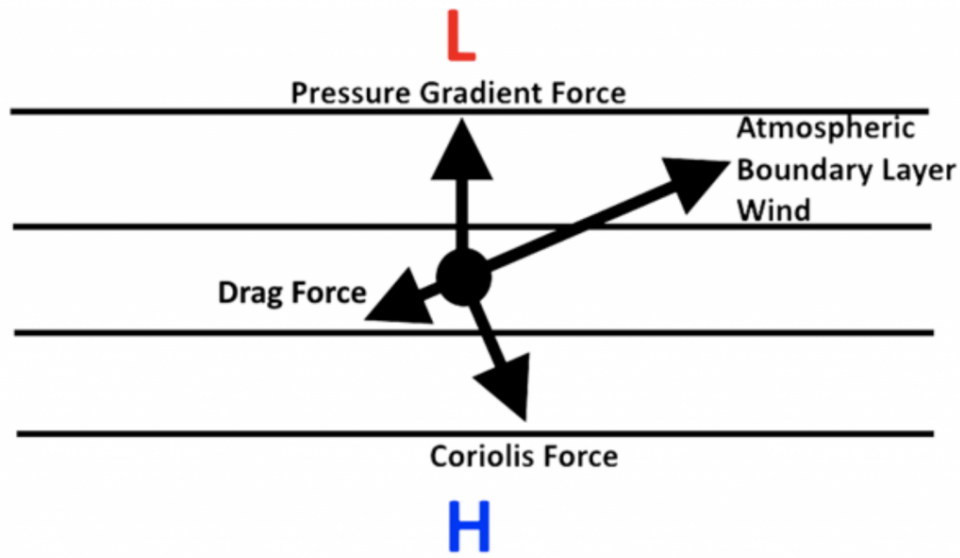



Figure 6.2.3: Atmospheric boundary layer or surface wind force balance includes the Pressure Gradient Force pulling the wind towards the low pressure, the Coriolis Force deflecting it to the right, and the Frictional Drag Force opposite to the direction of wind. As a result, surface wind angles towards the low-pressure system. (CC BY-SA 4.0; Shintaro Russell via Atmospheric Processes and Phenomena)

## Cyclostrophic Wind

Cyclostrophic wind occurs at smaller cyclonic scales (mesoscale), such as tornadoes, waterspouts, and even at the center of a

tropical cyclone. Because the scale is small, the Coriolis force is negligible. When a small cyclonic scale, such as a tornado, first forms, both the tangential winds and the centrifugal force increase much faster than the Coriolis force due to the very strong pressure gradient. As a result, the centrifugal force balances the pressure gradient force, as shown in Figure 6.2.4, with the Coriolis force neglected. 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. Anticyclones (highs), however, typically lack strong pressure gradients. Thus, winds around the high are too weak to be in cyclostrophic balance.

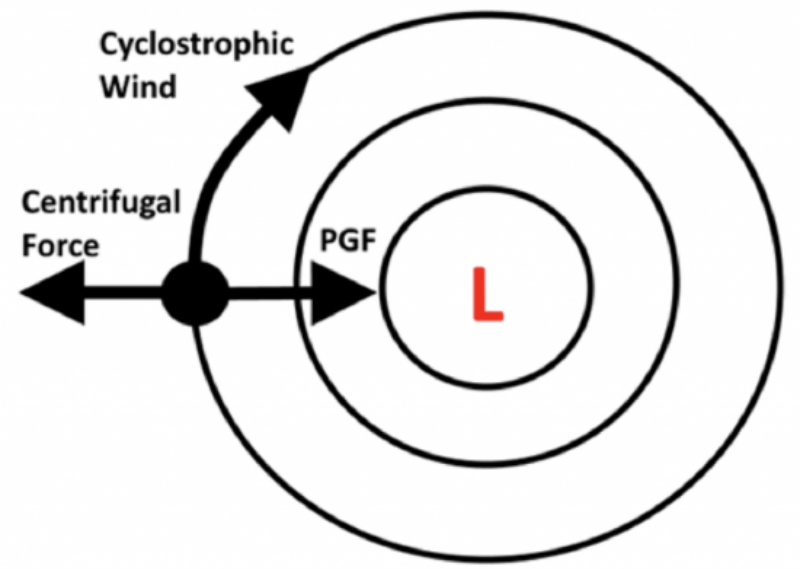



Figure 6.2.4: Cyclostrophic wind force diagram where the centrifugal force balances the pressure gradient force. The Pressure Gradient Force pulls the wind towards the low pressure, while the Centrifugal Force pulls it outward. If they balance each other, the wind blows at a constant speed around the low-pressure system. (CC BY-SA 4.0; Shintaro Russell via Atmospheric Processes and Phenomena)

All of the wind balances discussed (geostrophic balance, gradient wind, surface wind, and cyclostrophic wind) occur in Earth's atmosphere under differing conditions. The following chapters will clarify these applications.

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# 6.3: Exercises

## Chapter 6: Exercises

1. Explain why wind occurs.
2. Which forces influence the direction and speed of horizontal winds?
3. Match the Force Balances:

**Query 6.3.3**

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# 7: GENERAL CIRCULATION

## Introduction

Have you ever wondered why it rains at some times of year and not others? Or why some regions don't get much rain and are desert-like, while others get constant rain, resulting in lush rainforests (Figure 7.1)? While Chapter 5 showed how precipitation forms in general, we haven't yet covered its distribution worldwide. The primary reason for where (and when) precipitation falls is where storms travel, and that is dependent on wind circulations. The reason we have global wind patterns is ultimately due to a differentially heated, rotating Earth. The differential heating of Earth continually creates an imbalance in air pressure and temperature around the world, which, in turn, drives a continuous general circulation of winds that attempt to restore balance.



Figure 7.1: Rainforests such as this are located in areas where air converges and rises, producing immense stormy weather. (CC BY 4.0; Vyacheslav Argenberg via Wikimedia Commons)

While actual winds in a given place and time may differ from the average general circulation, the average can explain 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 the transport of heat and momentum from the equator to the poles.

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 6. The average global winds are called the **general circulation of the atmosphere**. To determine typical wind circulation, one needs to average wind speed and duration over a long period. Averaging over time removes short-duration fluctuations, allowing the primary sense of movement to be visualized.

### 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

7.1: Differential Heating
7.2: Single-Cell Model
7.3: Three-Cell Model
7.4: Jet streams
7.5: Exercises for Chapter 7
7.6: Accessible Descriptions

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# 7.1: Differential Heating

## Uneven Heating of Earth

Because the Earth is round, solar radiation is not evenly distributed across 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, as shown in Figure 7.1.1. 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, incoming solar radiation is distributed along latitude lines.

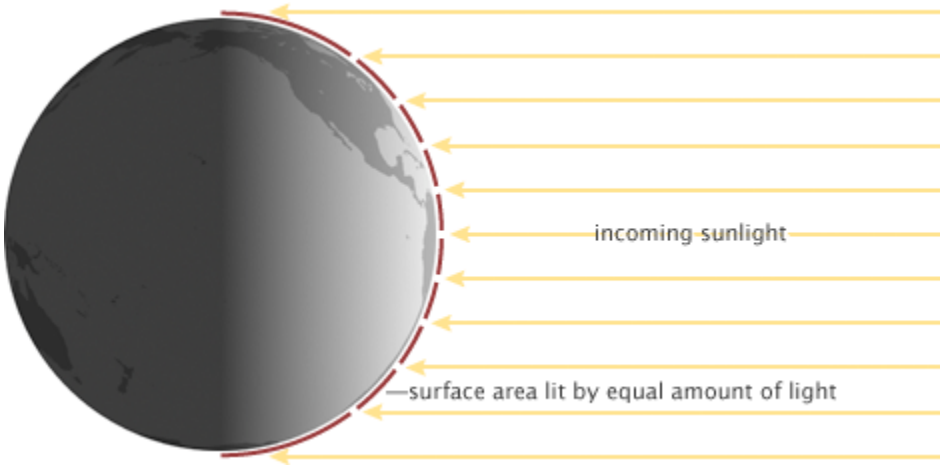



Figure 7.1.1: Earth's uneven heating by the sun due to the curvature of its surface. Incoming sunlight provides equal amounts of energy across parallel beams, while Earth's curved surface causes that energy to be spread over different surface areas. The illuminated portion of the planet receives the same total energy. Still, the curvature results in some regions receiving concentrated energy, while others receive the same energy distributed over a larger area. (Public Domain via NASA).

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, so lower latitudes are 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 constantly 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 Earth emits IR radiation according to the Stefan-Boltzmann law, which is highly dependent on temperature.

When averaged globally 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. The Earth receives more solar energy in the tropics, and while cooling by outgoing IR radiation helps offset this, there is still a net radiative energy gain 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.

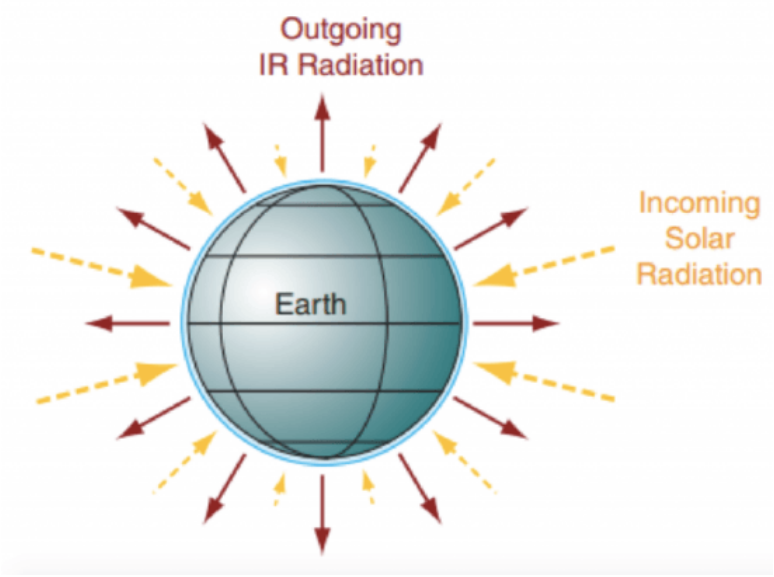



Figure 7.1.2: Incoming solar radiation (yellow dashed) and outgoing infrared radiation (red solid) diagram (CC-BY-NC-SA 4.0; Roland Stull via Practical Meteorology).

Figure 7.1.2 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 an energy surplus near the equator and a deficit toward the poles, as seen in Figure 7.1.3.

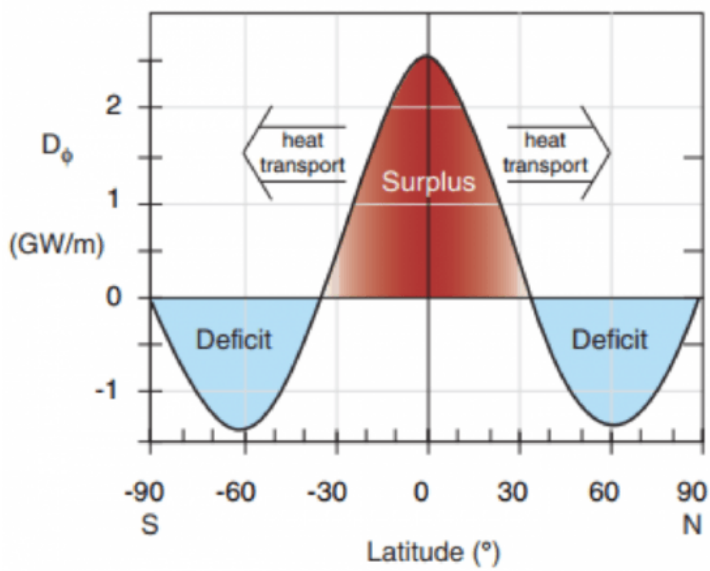



Figure 7.1.3: Earth gains more energy than it loses in low latitudes and loses more energy than it gains in high latitudes, creating an energy surplus near the equator and energy deficits toward the poles. The imbalance drives heat transport from surplus regions toward deficit regions. (CC-BY-NC-SA 4.0; Roland Stull via Practical Meteorology). Alternative Description of Image.

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. To build an understanding of this process, we'll start with simplified models and gradually increase complexity.

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# 7.2: Single-Cell Model

## Hadley Cells

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 be concerned only with the pressure gradient force.

With these assumptions in place, Earth's global circulation would be as illustrated in Figure 7.2.1, 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 warm air rises and cold air sinks.

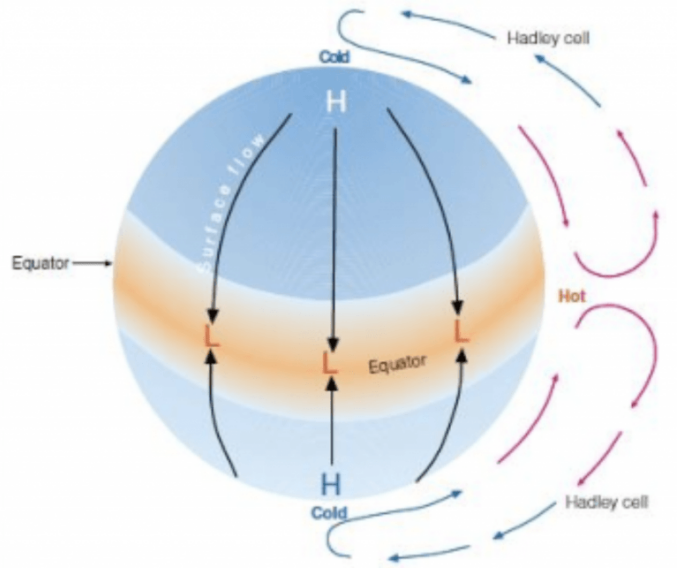



Figure 7.2.1: The single-cell model of Hadley cells on a planet. Warm air on the equator rises, generating low surface pressure, while cold air on the poles sinks, generating high surface pressure. At the surface, winds blow from the high pressure at the pole to the low pressure at the equator, while aloft, winds blow from the rising air at the equator to the sinking air at the poles. (CC-BY-SA 4.0; Suhari At Tamfury, via Wikimedia Commons).

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 excess heating of air at the equator creates a large area of low pressure at the planet's surface, while excess cooling at the poles creates a high-pressure area at the surface. This global horizontal pressure gradient causes air to flow from high to low at the surface (pole to equator), where it subsequently rises at the equator, flows back to the poles, and sinks.

Both reasons are plausible; it's a matter of whether you focus on temperature or pressure. The temperature and pressure differences are intertwined, and both are important for the general circulation. While this single-cell model can explain some phenomena and works in some ways (and on some planetary bodies), it is not the reality on Earth. Because Earth is rotating, 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 to 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 the Earth. A different model is needed.

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# 7.3: Three-Cell Model

## Adding Rotation to the 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 Figure 7.3.1. It may look more complex and less related to the single-cell model, but there are many similarities. There is still excess heating in equatorial regions and excess cooling in the polar areas. 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 low-pressure belt. Let's examine what happens to the rising air just above the equator.

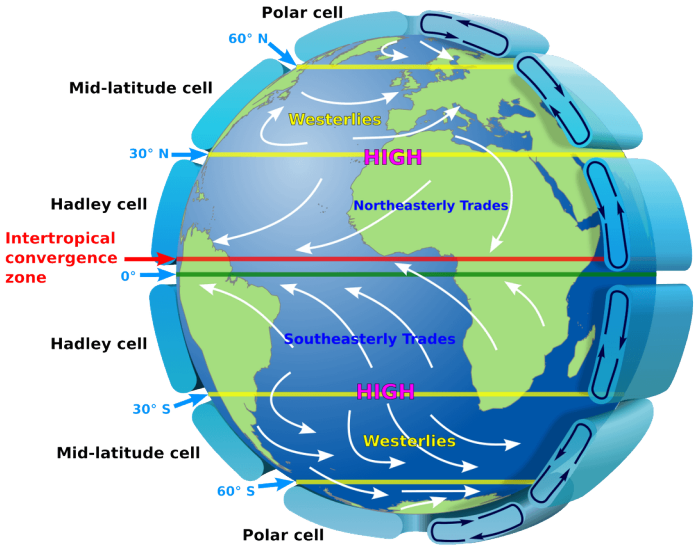



Figure 7.3.1: Three-cell model of the rotating Earth and the resulting wind circulations. Low pressure is present over the equator and 60°N/S, while high pressure is present at 30°N/S and the poles. There are three overturning circulations: The Hadley cell between the equator and 30°N/S, the Ferrel cell between 30°N/S and 60°N/S, and the Polar cell between 60°N/S and the poles. Trade winds between 30°N/S and the equator blow from the northeast in the Northern hemisphere wile they blow from the southeast in the Southern hemisphere. Westerly winds blow between 30°N/S and 60°N/S in both hemispheres. (CC-BY-SA 3.0; Kiador via Wikimedia Commons).

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 further upward motion, 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 aloft increases the mass above the surface, thereby raising surface pressure. 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**. Near the equator, the northeasterly and southeasterly trade winds converge at the surface, forming 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 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 the weather moves from 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 following 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 reaches the tropopause, where it moves back toward 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° latitude. As the air moves equatorward, it is again deflected by the Coriolis force. In the Arctic, air typically flows from the northeast, while in the Antarctic, air typically 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, where it is deflected into a westerly wind aloft. Eventually, this air reaches the poles, sinks back to the surface, and flows back toward the polar front, forming the **Polar cell**.

To summarize, in the three-cell model, 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 surface pressure, while regions where they diverge will have sinking motion and high surface pressure.

How does this three-cell model match with reality? While some minor discrepancies exist, for example, in fact, much of the upper-level winds in the mid-latitudes are westerly, like the surface. At the same time, the Ferrel cell suggests easterly winds aloft. However, this model is roughly accurate for surface winds and provides a really good first-order pattern for general circulation. How does the real world's average surface sea level pressure field compare with Figure 7.3.1? When we add in the continents, ice masses, oceans, mountains, and forests, we get an average that is represented by Figure 7.3.2 and Figure 7.3.3.

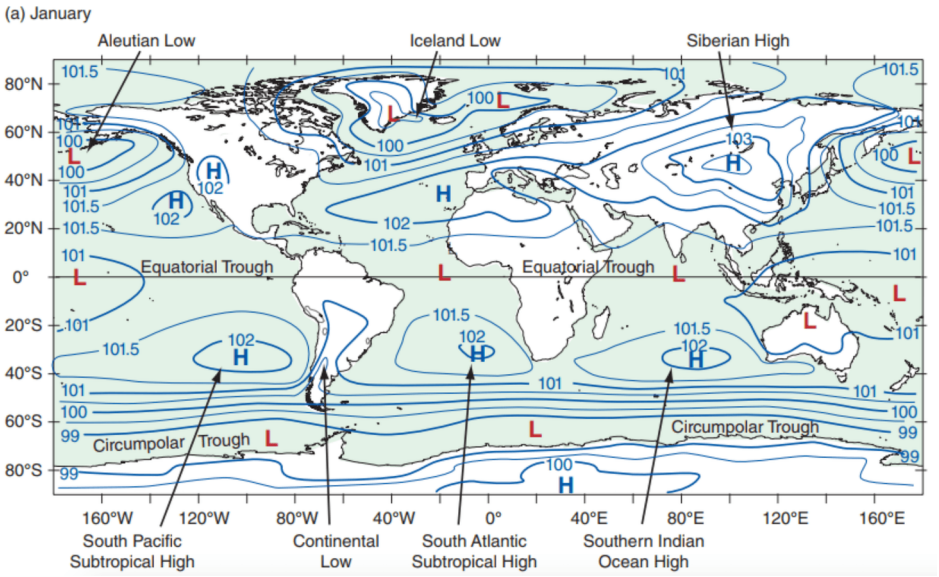



Figure 7.3.2: Map of semipermanent highs and lows during January. (CC BY-NC-SA 4.0; Roland Stull via Practical Meteorology). Alternative Description of Image.

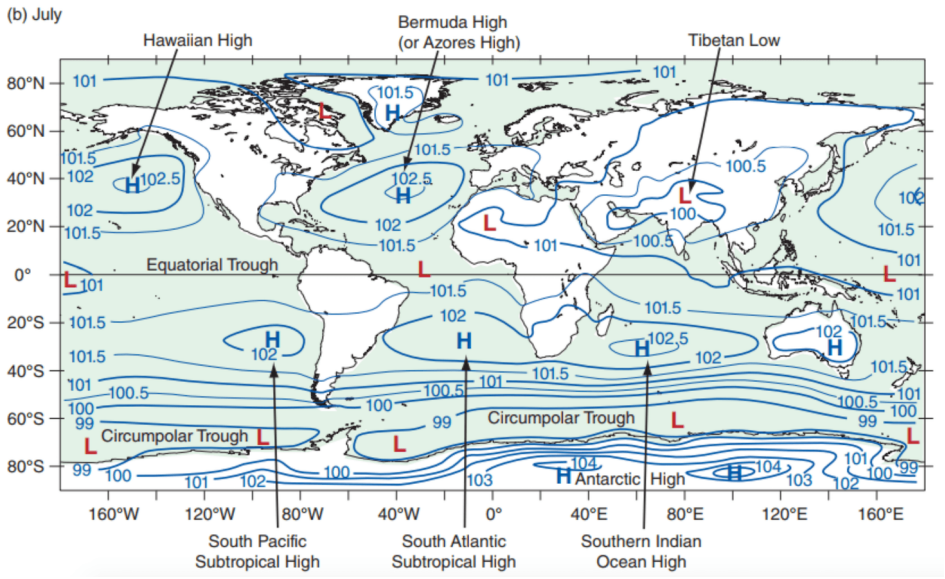



Figure 7.3.3: Map of semipermanent highs and lows during July (CC BY-NC-SA 4.0; Roland Stull via Practical Meteorology). Alternative Description of Image.

Figure 7.3.2 and Figure 7.3.3 show the mean sea-level pressure field for January and July, averaged from 1981 to 2010. 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**.

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# 7.4: Jet streams

## Polar and Subtropical Jet Streams

There is one final important piece of general circulation that deserves discussion: jet streams. In Figure 7.4.1 you can see two jet streams: the subtropical jet and the polar jet. The figure shows the average positions of the jet streams in the Northern Hemisphere during winter, along with their relationship 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 above the surface, at latitudes of 50° to 60 °. The difference in height of these jets is due to their location at the tropopause, and to the fact that the tropopause is higher in tropical regions than in polar regions, reflecting differences in the average temperature of the troposphere below. The troposphere is thinner in polar regions than in tropical regions because air is colder and denser at the poles.

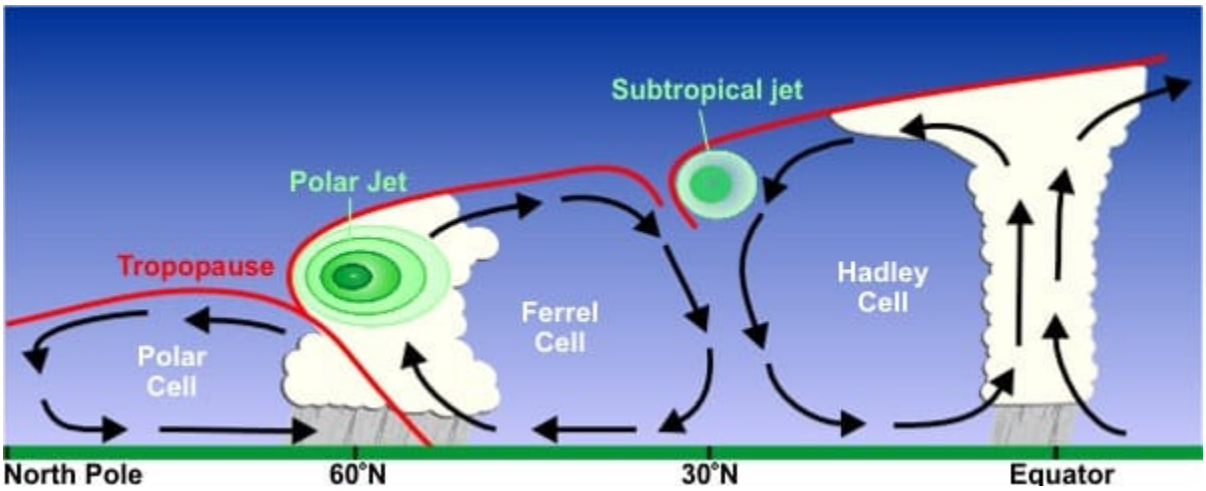



Figure 7.4.1: Cross-section of the northern hemisphere circulation, and the positions of the polar and subtropical jet streams. Details in the above paragraph (Public Domain; National Weather Service via Wikimedia Commons). Alternative Description of Image.

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.

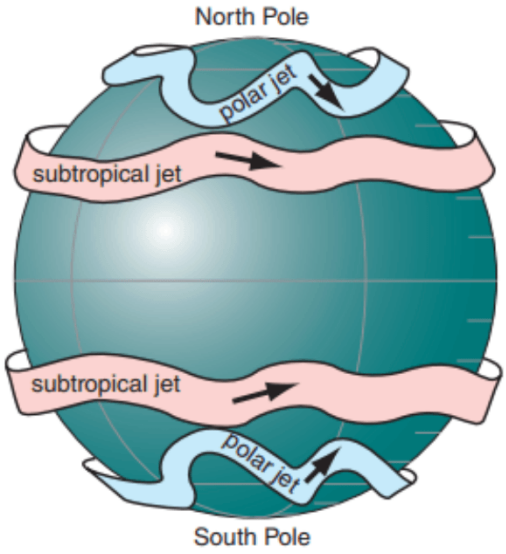



Figure 7.4.2: A global schematic highlighting the main jet streams in each hemisphere. Two subtropical jet streams encircle the planet in the midlatitudes, forming fast-moving bands linked to strong temperature contrasts between tropical and subtropical air. Closer to the poles, narrower polar jet streams mark the boundaries between cold polar air and warmer midlatitude air. The pattern appears in both hemispheres, illustrating how steep temperature gradients at different latitudes produce distinct high-altitude wind bands that strongly influence weather systems (CC BY-NC-SA 4.0; Roland Stull via Practical Meteorology).

The existence of jet streams ultimately results from the energy imbalance between the 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. As a result, the strongest temperature gradient occurs along the polar front. This rapid change in temperature with distance also causes a rapid change in pressure, 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 generates intense winds that form 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 extends farther south into subtropical regions. 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 called the *subtropical front*, but it does not extend to the surface. Here, the temperature gradient is strongest aloft near the tropopause, which induces a sharp pressure gradient and strong winds.

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# 7.5: Exercises for Chapter 7

## Chapter 7: Exercises

1. Interactive Content

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   View this Interactive Content online

2. What assumptions are made for a single-cell model of Earth's atmosphere?
3. What changes (from the assumptions in question 2) are made to the three-cell model?
4. Where are jet streams located? Why do they differ in height?

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# 7.6: Accessible Descriptions

## Figure 7.1.3: Energy Balances with Latitude

Graph showing the latitudinal distribution of Earth's net radiation imbalance and resulting heat transport. The horizontal axis represents latitude from 90° south to 90° north, and the vertical axis represents net energy (in gigawatts per meter). The curve is positive in low latitudes near the equator, indicating a surplus of energy, and negative in higher latitudes toward both poles, indicating a deficit. The surplus region peaks at the equator and transitions to deficit around mid-latitudes. Arrows indicate that heat is transported from the equatorial surplus regions toward higher latitudes in both hemispheres. Approximate net radiation values for various latitudes are provided in Table 7.6.1.

*Table 7.6.1: Approximate net radiation values for various latitudes.*

| Latitude (°) | Net radiation (GW/m) | Region classification |
|---|---|---|
| 90 S | -1.3 | Deficit |
| 60 S | -1.5 | Deficit (minimum) |
| 45 S | -0.8 | Deficit |
| 30 S | 0.0 | Transition |
| 15 S | 1.5 | Surplus |
| 0 | 2.5 | Maximum surplus |
| 15 N | 1.5 | Surplus |
| 30 N | 0.0 | Transition |
| 45 N | -0.8 | Deficit |
| 60 N | -1.5 | Deficit (minimum) |
| 90 N | -1.3 | Deficit |

## Figure 7.3.2: Average Position of Pressure Systems in January

Global map showing average sea-level pressure patterns for January, with contour lines labeled in hectopascals (hPa). In the Northern Hemisphere, a low-pressure center over the North Pacific (Aleutian Low) is shown with values around 1000 hPa near 50–60°N, and another low over the North Atlantic (Icelandic Low) also near 1000 hPa around 60–70°N. A strong high-pressure system over Asia (Siberian High) reaches about 1030 hPa between roughly 40–60°N. Subtropical high-pressure systems appear in the Southern Hemisphere, including the South Pacific, South Atlantic, and southern Indian Ocean highs, each around 1020 hPa near 30–40°S.

Near the equator (0–10° latitude), a band of lower pressure called the equatorial trough is indicated, with values around 1008–1010 hPa. In the Southern Hemisphere mid- to high latitudes (about 50–70°S), a continuous belt of low pressure known as the circumpolar trough is shown, with values generally between about 990 and 1000 hPa. Additional continental low-pressure regions in tropical land areas are marked with pressures around 1005–1010 hPa.

Overall, the map shows higher pressures in the subtropics (~1015–1025 hPa), lower pressures near the equator (~1008–1012 hPa), and pronounced low-pressure systems in subpolar regions (~990–1000 hPa), with strong seasonal contrasts in the Northern Hemisphere during winter.

## Figure 7.3.3: Average Position of Pressure Systems in July

Global map showing average sea-level pressure patterns for July, with contour lines labeled in hectopascals (hPa). In the Northern Hemisphere, high-pressure systems dominate over the oceans, including the Hawaiian High in the North Pacific (around

1020–1025 hPa near 30–40°N) and the Bermuda or Azores High in the North Atlantic (also about 1020–1025 hPa near 30–40°N). Over Asia, a strong low-pressure system known as the Tibetan Low is centered around 1000 hPa near 20–30°N.

Near the equator (0–10° latitude), the equatorial trough is present with pressures around 1008–1010 hPa. In the Southern Hemisphere, subtropical high-pressure systems are located over the South Pacific, South Atlantic, and southern Indian Ocean, each around 1020–1025 hPa near 30–40°S. Farther south, a continuous belt of low pressure called the circumpolar trough occurs between about 50–70°S, with pressures around 990–1000 hPa. Over Antarctica, a strong high-pressure system (Antarctic High) reaches about 1030–1040 hPa near 80–90°S.

Overall, the map shows stronger oceanic high-pressure systems in the Northern Hemisphere during summer, a continental low over Asia, and persistent Southern Hemisphere pressure belts.

## Figure 7.4.1: Three Circulation Cells and Jet Streams

Diagram showing large-scale atmospheric circulation in the Northern Hemisphere from the equator to the North Pole. The horizontal axis represents latitude, labeled at the equator (0°), about 30°N, about 60°N, and the North Pole (90°N). Three circulation cells are depicted. From the equator to about 30°N, the Hadley cell shows rising air at the equator, poleward flow aloft, sinking air near 30°N, and return flow toward the equator at the surface. Between about 30°N and 60°N, the Ferrel cell shows surface flow toward higher latitudes, rising motion near 60°N, and return flow aloft toward 30°N. From about 60°N to the pole, the Polar cell shows rising air near 60°N, poleward flow aloft, sinking air over the pole, and surface flow back toward 60°N.

Two jet streams are indicated: the subtropical jet near 30°N and the polar jet near 60°N, both located near the tropopause. The tropopause is shown as a boundary that slopes downward from the equator toward higher latitudes. The diagram emphasizes that air rises at approximately 0° and 60°N and sinks near 30°N and 90°N, with jet streams forming at the boundaries between circulation cells.

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# 8: FRONTS AND AIRMASSES

## Introduction

We've already learned that the Sun heats the Earth unevenly. This results in some areas with low pressure and others with higher pressures. Low- and high-pressure regions can be identified on a weather map by the isobars. These pressure differences drive winds from the high-pressure areas towards low-pressure regions. When this happens, air from different pressure and temperature regions on the surface collides, as illustrated in Figure 8.1.

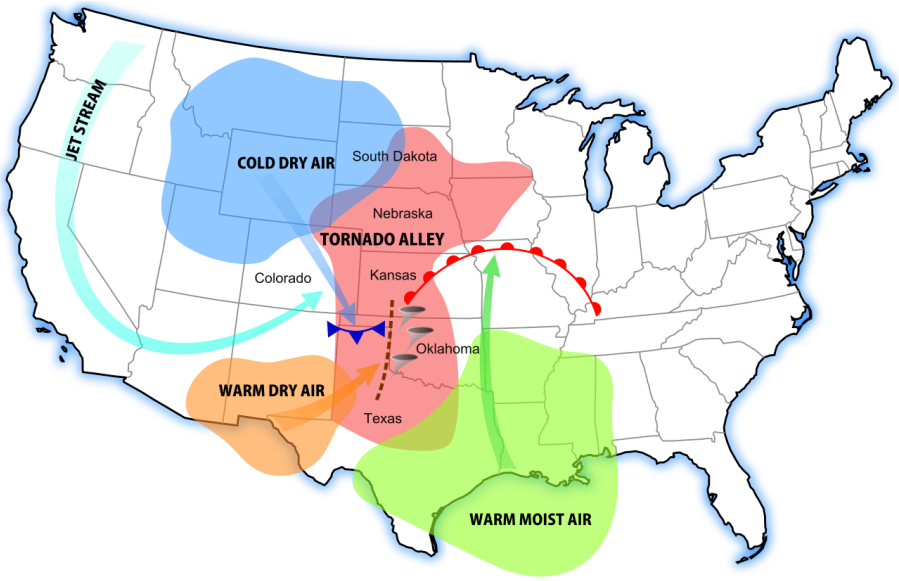



Figure 8.1: An interaction between different fronts and air masses. A cold front extends southward, with a warm front heading northward. Warm air is located ahead of the cold front, and warm, moist air is entering behind the warm front, while cool air exists ahead of the warm front, and cold air is present behind the cold front. The interaction between the different air masses and fronts leads to tornado formation in Tornado Alley. (CC BY-SA 3.0; Dan Craggs via Wikimedia Commons)

In this chapter, we'll learn that depending on temperature and moisture, some regions of the atmosphere have similar air properties and are named by those properties as a collective mass of air or "air mass". The interaction between different air masses significantly influences local weather.

### Learning Objectives

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

- Describe what an "air mass" is and name some of the main types.
- 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.

8.1: Air Masses
8.2: Fronts
8.3: Mid-latitude Frontal Cyclones
8.4: Cyclogenesis

8.5: Exercises

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# 8.1: Air Masses

## Types of 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 by their temperature and humidity characteristics, as well as their geographical regions of origin. 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). 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.

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 consist of extremely cold air originating 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). Figure 8.1.1 illustrates where these different types of air masses typically originate.

- 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.

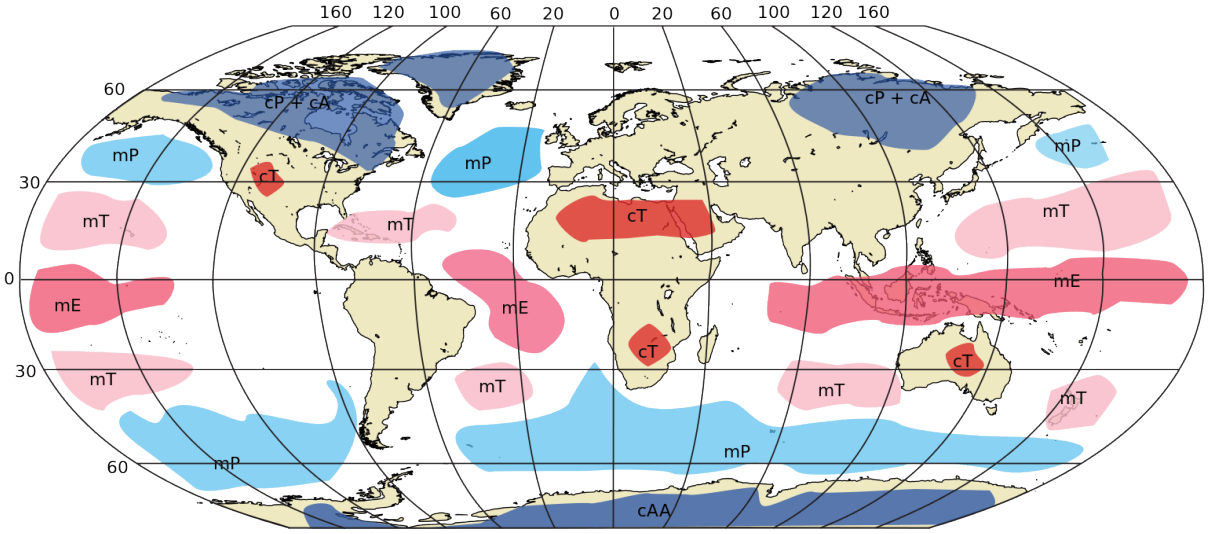



Figure 8.1.1: Temperature and pressure vary over different regions of the world and result in different air masses in each region. Based on temperature and moisture, each air mass is assigned an abbreviated name that identifies its origin. Maritime air masses form over the oceans, while continental air masses form over land. Tropical and Equatorial air masses are found in the tropics, while polar air masses occur over the polar latitudes. (Public Domain; David Hall via Wikimedia Commons).

## Creation of Air Masses

Air masses develop when air remains over a surface for an *extended period*. 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 the stable air over cold surfaces experiences less turbulence and thus less mixing (while greater turbulence over warm surfaces creates more mixing). As air masses shift from their source regions, they change over time as they pass through the surfaces and terrain through which they 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 different surface characteristics, the air mass begins to change. 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 typically dries as it crosses the mountains, releasing 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 two air masses collide, they form a boundary known as a **front**.

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# 8.2: Fronts

## Surface Fronts

Surface fronts are the boundaries or transition zones between air masses at the Earth's surface. Changes in temperature, humidity, wind, pressure, and visibility, as well as specific 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

Cold fronts are transition zones in which the advancing cold air mass replaces the retreating warm air mass. On weather maps, cold fronts are depicted as blue lines with blue triangles pointing toward the warmer air mass, indicating the direction of frontal movement, as illustrated in Figure 8.2.1.

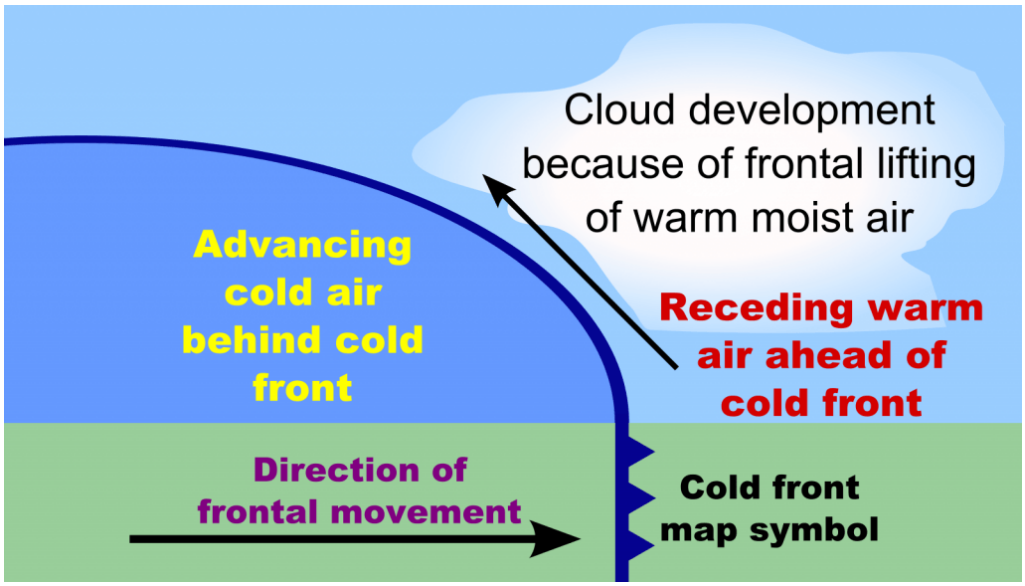



Figure 8.2.1: A vertical cross-section of a cold front shows cold air behind the cold front (dark blue lines) advancing into warmer air ahead of the front. Where the two air masses meet, convergence often occurs, resulting in the upward motion of air parcels. If the air contains enough moisture, rain can occur. (CC BY-SA 4.0; Ravedave via Wikimedia Commons).

Because warm air is less dense than cold air, the cold air stays at the bottom, and warm air is forced to rise above the advancing cold air. This forced lifting produces typical cloud patterns ahead of a cold front, including cirrus and cirrostratus. 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

Warm fronts are transition zones in which the advancing warm air mass replaces the retreating cold 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, as illustrated in Figure 8.2.2.

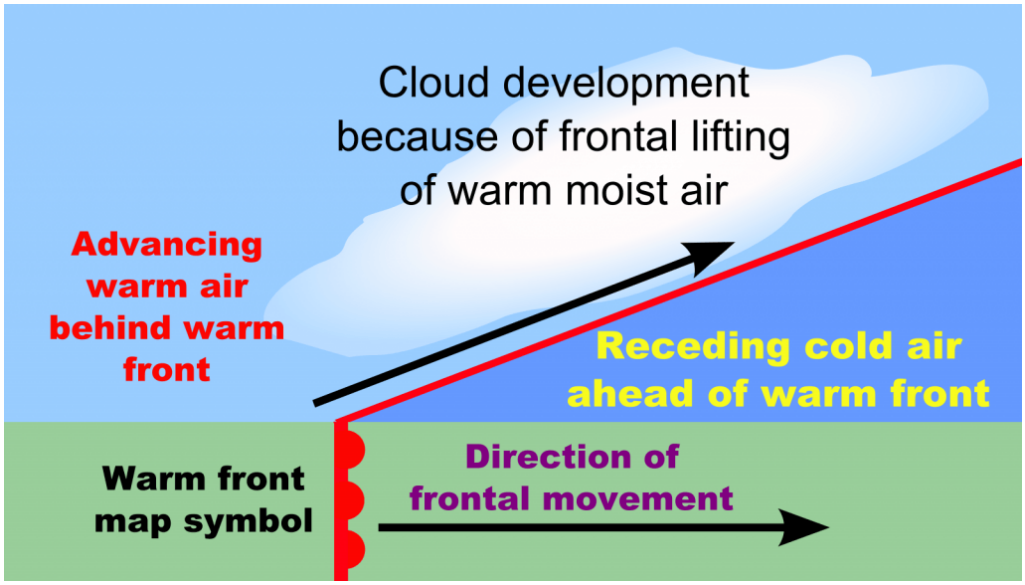



Figure 8.2.2: A vertical cross-section of a warm front (dark red lines) shows how surface warm air behind the warm front flows upward in a slantwise manner over top low-level cool or cold air ahead of the position of the surface warm front. This causes clouds and precipitation ahead of the surface warm front when the air is sufficiently moist. (CC BY-SA 3.0; Ravedave via Wikimedia Commons).

Warm air advancing over the retreating cold, dense air is forced to rise. 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, before clouds thicken, and lower-level clouds, such as altostratus, nimbostratus, and fog, near the frontal boundary. As the air mass rises 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 decrease, with only scattered cumulus clouds remaining.

## Occluded Fronts

Occluded fronts are a frontal boundary that forms when a cold front catches up to a warm front, as illustrated in Figure 8.2.3. Cold fronts move faster than warm fronts, so cold fronts can often catch up to warm fronts, but not the other way around. When this happens, the warm air is separated (occluded) from the cyclone center at the surface. The cold air from the cold front and cooler air ahead of the warm front cover the surface near the front, while warm, moist air is convected to higher elevations. The point where the warm front becomes the occluded front is called the *triple point*.

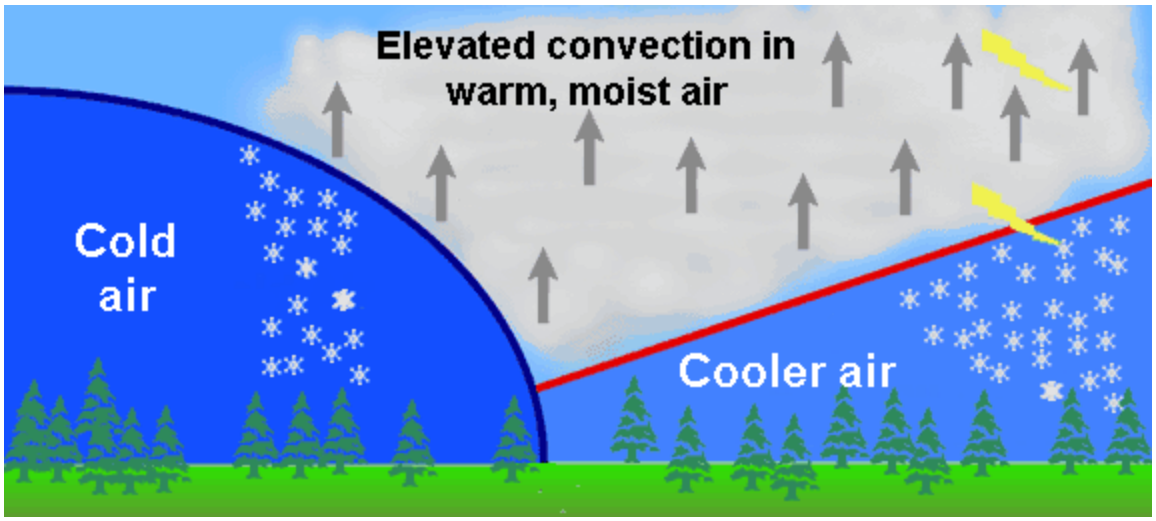



Figure 8.2.3: A vertical cross-section through an occluded front shows that the cold air mass is moving faster than the cool air mass ahead. As the two fronts converge, the cold air undercuts the warmer air mass. (Public Domain; JWIZMAN96 via Wikimedia Commons).

There are two types of occlusions: cold occlusions and warm occlusions. Cold occlusions, as illustrated in Figure 8.2.3, 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. Warm air occluded fronts are more common on the West Coast of continents, while cold air occluded fronts are more common on the East Coast of continents. On weather maps, occluded fronts are drawn as purple lines with purple triangles and purple semicircles, as illustrated in Figure 8.2.4. 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. Occluded fronts are the final stage of a frontal boundary because the warm air above the cool air is stable.



Figure 8.2.4: Frontal symbol for an occluded frontal boundary (Public Domain; -xfi- via Wikimedia Commons).

## Stationary Fronts

Stationary fronts are a type of frontal system in which the winds flow nearly parallel to one another, from opposite directions on each side, separated by the front. On weather maps, stationary fronts are depicted as alternating blue and red lines, with blue triangles indicating the direction of the warmer air mass and red semicircles indicating the direction of the colder air mass, as illustrated in Figure 8.2.5. This is the only scenario in which the direction of the symbols does not indicate movement.

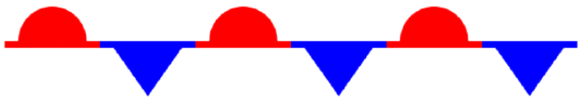


Figure 8.2.5: Frontal symbol for a stationary frontal boundary (Public Domain; -xfi- via Wikimedia Commons).

## Other Frontal-like Features

Not all weather features have frontal characteristics, such as a trough and a dryline. While troughs feature changes in cloud and precipitation patterns, they lack the sharp temperature and moisture gradients 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 the 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.

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# 8.3: Mid-latitude Frontal Cyclones

## Cyclogenesis and Life Cycle

Early weather forecasters observed a pattern that occurred whenever significant precipitation occurred in the Northern Hemisphere mid-latitudes. First, pressure would begin to drop, then the winds would increase, then precipitation would occur, and finally the skies would clear again. The wind direction would also change in a predictable manner, similar to a tropical cyclone. Based on this pattern, meteorologists in Bergen, Norway, developed a model to explain the life cycle of a "mid-latitude cyclone". According to this Norwegian model, a mid-latitude cyclone develops along the polar front. The polar front is a stationary front that separates cold polar air from warmer subtropical air at around 60° latitude. It is a semi-continuous boundary, along which mid-latitude cyclones form and move as a series of waves. For this reason, a developing storm is sometimes referred to as a wave cyclone.

Figure 8.3.1 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 can set up a rotational motion along the front. This is similar to how a pen turns when you place it between your palms and move your left and right hands in opposite directions.

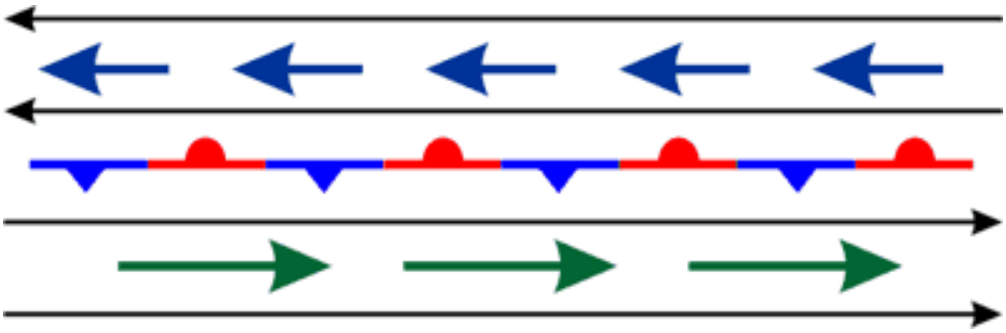


Figure 8.3.1: Part 1 of cyclogenesis consists of a stationary front with opposite-moving winds on either side. Cooler winds are to the north heading west, while warmer winds are to the south heading east. (Public Domain; Pflatau via Wikimedia Commons).

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 develops into a fully mature cyclone 12 to 24 hours after its incipient stage.

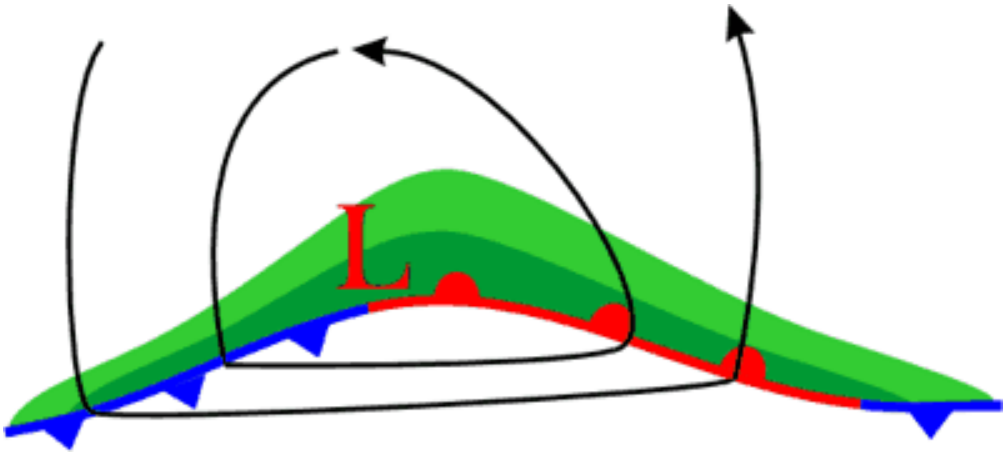



Figure 8.3.2: Part 2 of cyclogenesis includes the formation of a frontal wave or a frontal boundary in front of the low-pressure region. (Public Domain; Pflatau via Wikimedia Commons).

The central pressure lowers and the pressure gradient increases, causing a stronger cyclonic (counterclockwise) flow inward toward the low's center, as illustrated in Figure 8.3.2. 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 warmer and colder air masses attempt to regain equilibrium, warm air rises over colder air, converting potential energy into kinetic energy (motion). As warmer air rises, it condenses into clouds, releasing latent heat 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, as illustrated in Figure 8.3.3.

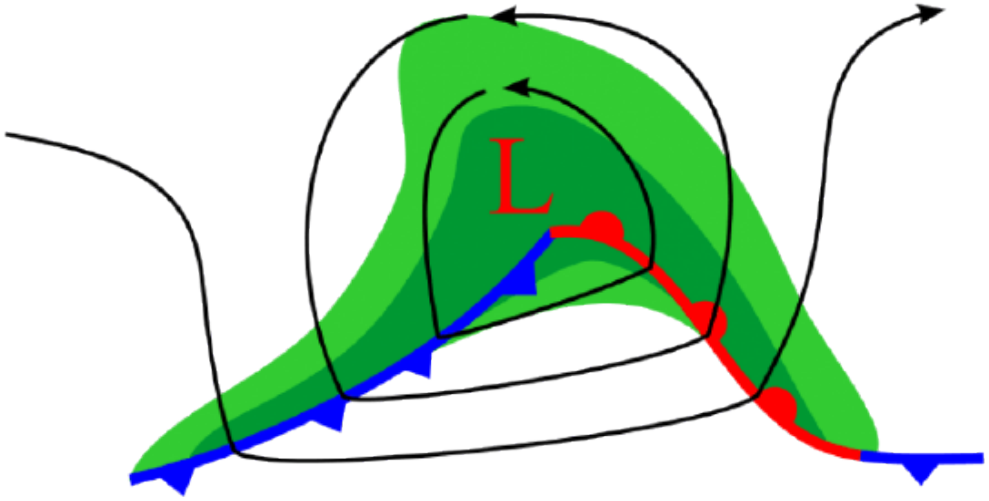



Figure 8.3.3: Part 3 of cyclogenesis results in a newly developed cyclone with a significantly lower pressure in the center, surrounded by a counter-clockwise wind pattern. (Public Domain; Pflatau via Wikimedia Commons).

As the cyclone moves eastward, the central pressure continues to decrease, and winds increase during its mature stage. As the faster-moving cold front advances toward the warm front, it reduces the warm sector's size. In the Norwegian cyclone model, the cold front overtakes the warm front, and the cyclone becomes occluded, as illustrated in Figure 8.3.4. Cloud cover and precipitation extend over a wide area, and the storm is typically most intense during this stage.

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

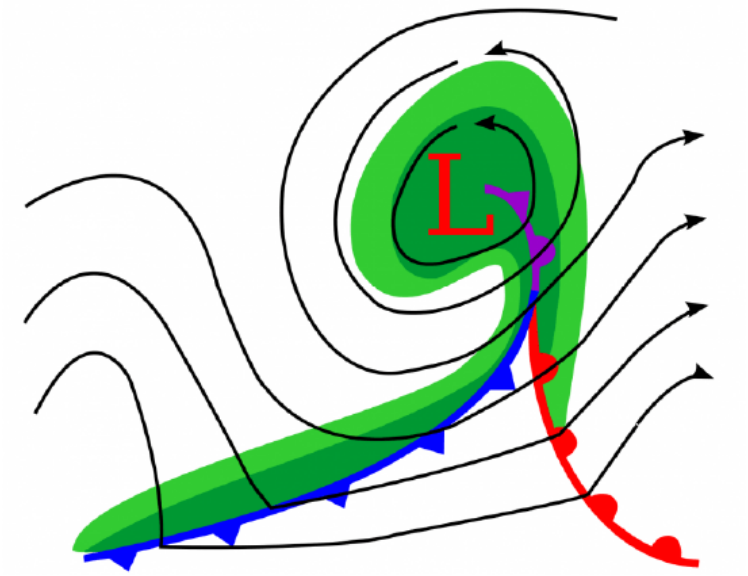



Figure 8.3.4: Part 4 of cyclogenesis results in the formation of an occluded front and "triple point" where the cold, warm, and occluded fronts intersect. (Public Domain; Pflatau via Wikimedia Commons).

Eventually, as occlusion advances, the low-pressure center will begin to dissipate because cold air exists on both sides of the occluded front, as illustrated in Figure 8.3.5. The warm, rising air is removed from the center of the storm, thereby cutting it off from its primary energy source. Eventually, the old storm dissipates and gradually fades away.

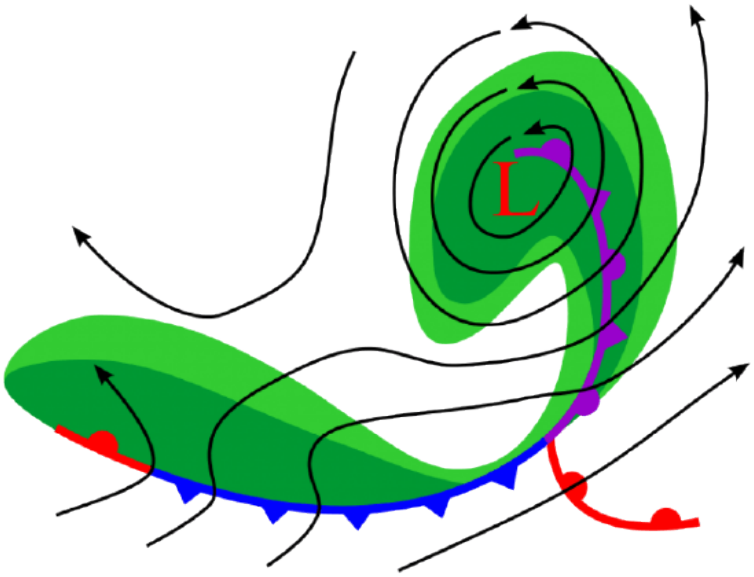



Figure 8.3.5: Part 5 of cyclogenesis includes the continued formation of an occluded front and the eventual dying out of the cyclone. (Public Domain; Pflatau via Wikimedia Commons).

This sequence of a developing mid-latitude cyclone is similar to a whirling eddy in a river that forms behind a stick or log, moves along with the river, and quickly disappears downstream. This entire life cycle can last from several days to a little more than a week. Cyclones at various stages of development can be observed simultaneously 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 somewhat simplified; in fact, very few storms follow this model exactly. However, it is a good foundation for understanding storm structure.

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# 8.4: Cyclogenesis

## Storm Tracks

The development of a mid-latitude cyclone is known as **cyclogenesis**. Certain regions in North America are more favorable for cyclogenesis, including the eastern slopes of mountain ranges such as the Rockies and the 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 contrast in air-mass temperature and moisture to the point that 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, as well as 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 redevelop on the leeward 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? Some surface conditions influence cyclogenesis, but the primary driver of mid-latitude cyclone development lies in the winds aloft. How does upper-level flow influence mid-latitude cyclones?

## 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 lies to the west of the surface low in the Northern Hemisphere. This is a necessary condition for a low-pressure system to continue developing and intensifying. 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 accumulates at the surface must have an "exit path" out of the column, allowing the surface air pressure to continue decreasing and the cyclone to strengthen. The vertical structure of the atmosphere must allow air to rise from a low-pressure system at the surface.

Figure 8.4.1 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.

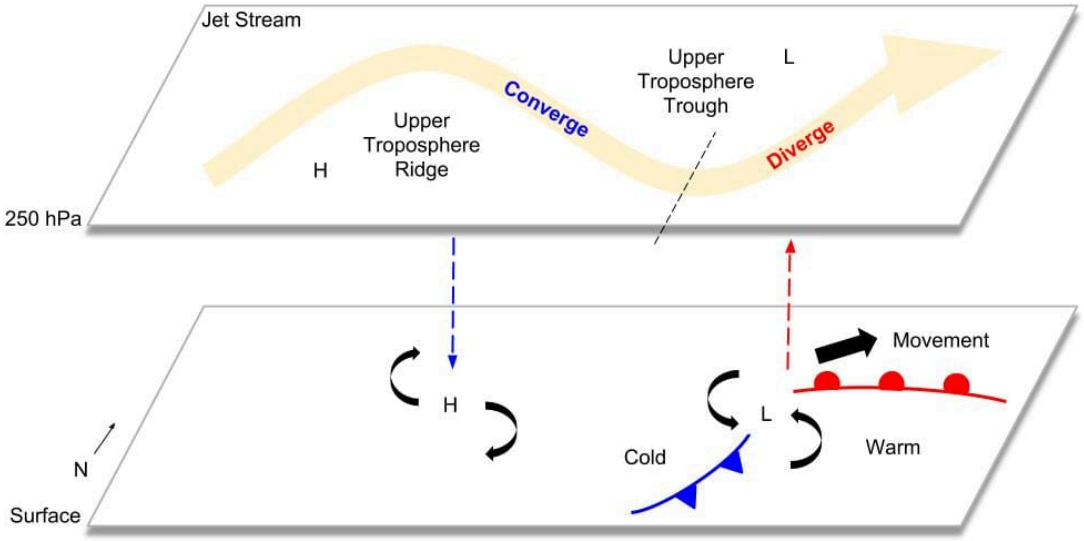



Figure 8.4.1: This diagram shows how the jet stream influences surface weather, with upper-level convergence over a ridge linked to surface high pressure, and upper-level divergence over a trough linked to rising air, low pressure, and warm and cold fronts. Air converges on the upstream side of an upper-level trough and diverges on the downstream side. A ridge is associated with a high-pressure system on the surface, and a trough is associated with a low-pressure system. (CC BY-SA 4.0; Britt Seifert via Atmospheric Processes and Phenomena).

On the right-hand side is a Northern Hemisphere frontal cyclone exhibiting both a warm and a cold front. The cold air behind the cold front at the surface also extends upward into the air 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 more closely spaced, pressure decreases more rapidly with height in a column of cold air. Because of this, low pressure is found in a body of cold air aloft, 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 as it moves upward. 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 divergence in the upper levels exceeds convergence at the surface, surface pressures will be lower farther out, 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.

Note the 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 must be continually added to the anticyclone. This happens when there is convergence above a surface high. The air that accumulates aloft sinks within 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 travel at approximately 16 knots in summer and roughly 27 knots in winter. This is due to stronger jets and upper-level flow in winter, resulting from greater north-south temperature contrasts.

## Redistribution of Heat

Mid-latitude frontal cyclones are both a vital part of the global circulation and a result of it. They're also an important pattern in the climatology of mid-latitude regions. 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, characterized by cold air from the polar region and warm air from the tropics, is the energy source that drives frontal storms. On the other hand, frontal cyclones are a significant contributor to the global redistribution of heat. 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 chapter began with images of these storm systems extending across scales comparable to continents and landmasses. It also described how these storm systems last from days to over a week. Instabilities along the polar front are always growing and dying as they pass over fixed points on Earth. The point is that if you live somewhere along the storm track in the Northern or Southern Hemisphere during wintertime, these storm systems dictate your weather. For example, in Boston, Massachusetts, winter weather may look something like this: a few days of warm, clear weather, followed by a sudden change (a passing cold front) that results in a significant drop in temperatures and some heavy rain, then cold, dry weather for a few days. This cold weather gradually transitions to warmer conditions, accompanied by light rain and rising temperatures (warm front). Although variable, this pattern recurs 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 wintry weather. Precipitation and temperature variations associated with frontal cyclones are a crucial component of the climatology of mid-latitude weather.

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# 8.5: Exercises

## Chapter 8: Exercises

1. What is an air mass? What are some different types of air masses, and where do they originate?
2. Which front do mid-latitude cyclones form and move along?
3. How long does it take for a cyclone to fully develop?
4. What is the point where the cold front, warm front, and occluded front intersect called?
5. What cloud and precipitation patterns are typically associated with warm and cold frontal passages?

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# 9: WEATHER REPORTS AND MAP ANALYSIS

## Introduction

The weather is the state of the atmosphere at a given time and is constantly changing. 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?

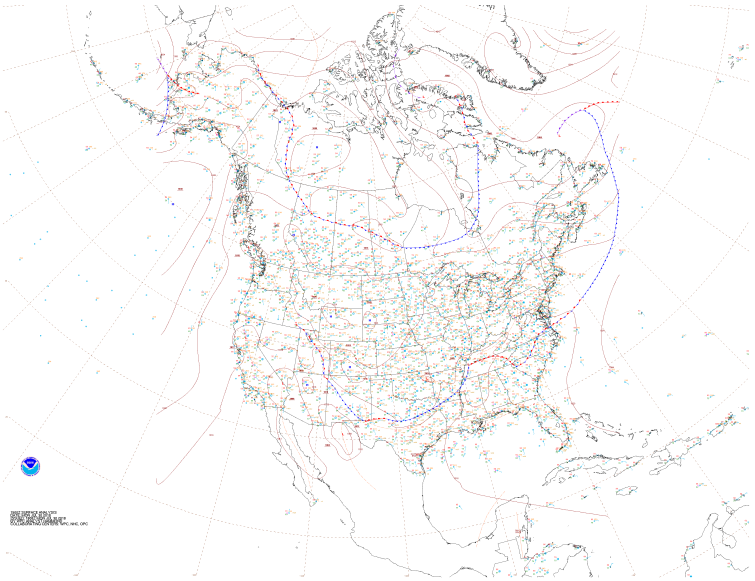


Figure 9.1: What a high-resolution surface weather map looks like. (Public Domain, North American Surface Analysis via NOAA).

This chapter serves as a brief introduction to surface synoptic weather maps and interpreting the data plotted on them. Simplified weather charts are frequently used on TV to show the locations of high- and low-pressure systems, fronts, and storm systems. You're likely familiar with those, just from your day-to-day experiences. But the high-resolution weather charts that meteorologists commonly use can be highly detailed and very confusing. A sample high-resolution surface weather map is shown in Figure 9.1.

### 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


9.1: Meteorological Reports and Observations

9.2: Synoptic Weather Maps

9.3: Map Analysis

9.4: Exercises for Chapter 9



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# 9.1: Meteorological Reports and Observations

## Reading METARs

The **World Meteorological Organization (WMO)** is a branch of the United Nations, and it sets global weather observation standards for countries worldwide. Weather systems span countries and continents, so weather observations must 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 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 are compiled from 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, and weather conditions such as precipitation or thunderstorms near the station. **SPECI** is a special non-routine METAR issued 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 the METAR code can be easily found online.

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

This example is a METAR from Honolulu International Airport. It is decoded in the steps below:

1. “PHNL” is the ICAO airport code for Honolulu International.
2. 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.
3. The “05012KT” wind report indicates that winds are blowing at 12 knots from 50°, roughly from the northeast.
4. The “10SM” indicates that visibility is 10 statute miles, which means the runway is clear. If visibility is 10 statute miles or greater, it is always reported as just “10SM”.
5. The “FEW025 FEW050 BKN200” are reports of different cloud heights, which are important for aviation purposes. This report says that there are is a layer of a “few” clouds at 2,500 and 5,000 feet, with a layer of “broken” clouds at 20,000 feet.
6. The “31/17” indicates that the temperature is 31°C with a dew point of 17°C. If there is an "M" located next to the temperature and/or dew point number, it indicates a "minus" sign. For example: 03/M02 would indicate an air temperature of 3°C and a dew point temperature of -2°C.
7. The “A3006” gives the station air pressure at 30.06 in Hg (inches of Mercury, this unit is still primarily used in aviation).
8. The “RMK” denotes the remarks section of a METAR report where additional remarks and information about the weather are provided.
9. The “AO2” code indicates that the airport observing site is automated and equipped with 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 include non-automated remarks.
10. The “SLP180” gives sea-level pressure as 1018.0 mb.
11. And the “T03060167” provides more accurate temperature and dew-point readings. It reports that the temperature is 30.6°C and the dew point is 16.7°C.

## Weather Observation Locations

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

Although surface weather observations are important, they tell only part of the 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 occurring at or near the ground is strongly influenced by conditions aloft.

Observations of the upper atmosphere are taken by **radiosonde** packages, which are carried aloft by weather balloons. These radiosonde observations (**RAOBs**) provide upper-air soundings, providing information on 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 measure 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 collected, organized, tested, and stored on computers at government centers such as the **European Center for Medium-Range Weather Forecasts (ECMWF)**. Other data-collection systems include weather radar (such as the **NEXRAD** network in the United States), satellite data, vertical **wind profilers,** and Radio Acoustic Sounding Systems (**RASS**).

The following figures show the locations of ECMWF data collected over a 6-hour period. Figure 9.1.1 shows surface observation locations located on land, Figure 9.1.2 shows surface observations over the ocean, and Figure 9.1.3 shows the upper air sounding observational network.

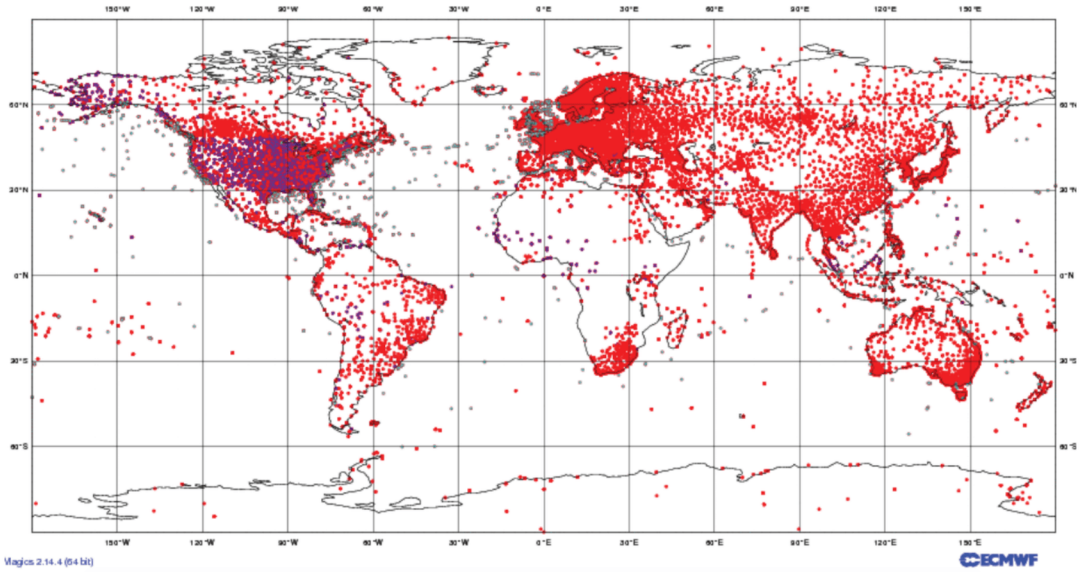



Figure 9.1.1: Surface observation locations for temperature, humidity, winds, clouds, precipitation, pressure, and visibility. The markers are densely clustered across major land regions, including North America, Europe, Asia, Australia, and parts of South America and Africa, with fewer points over remote land areas and oceans. The overall pattern indicates widespread global weather station coverage with higher concentrations in regions associated with greater data availability or monitoring activity. (CC-BY-NC-SA 4.0; Roland Stull, via Practical Meteorology).

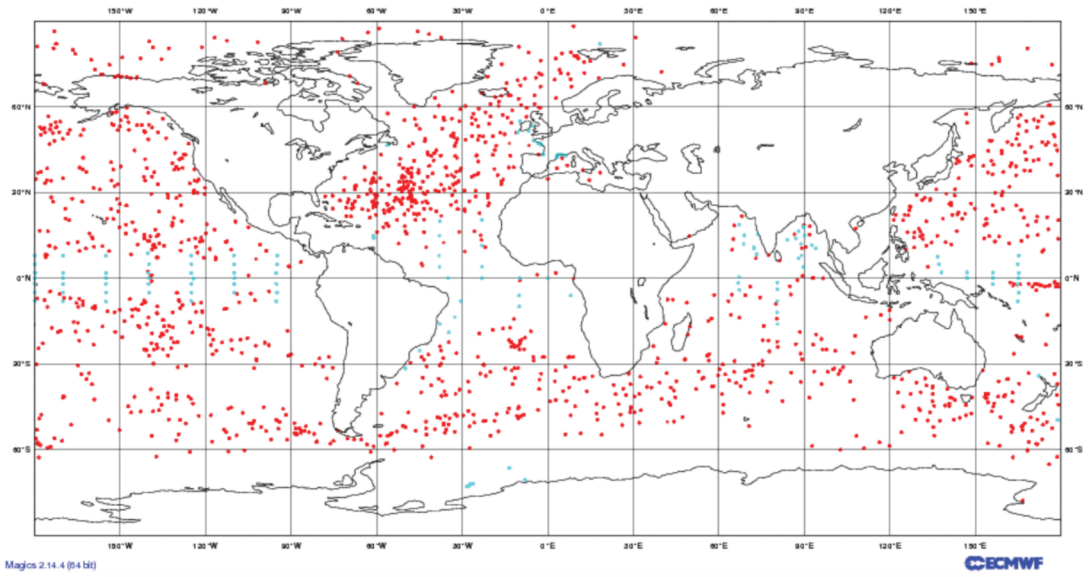



Figure 9.1.2: Buoy and ship-based observation platforms for surface temperature and winds over the ocean. The pattern indicates a global network of atmospheric data sources used for weather analysis or forecasting, with denser coverage where observation infrastructure or reporting activity is more frequent. (CC-BY-NC-SA 4.0; Roland Stull, via Practical Meteorology).

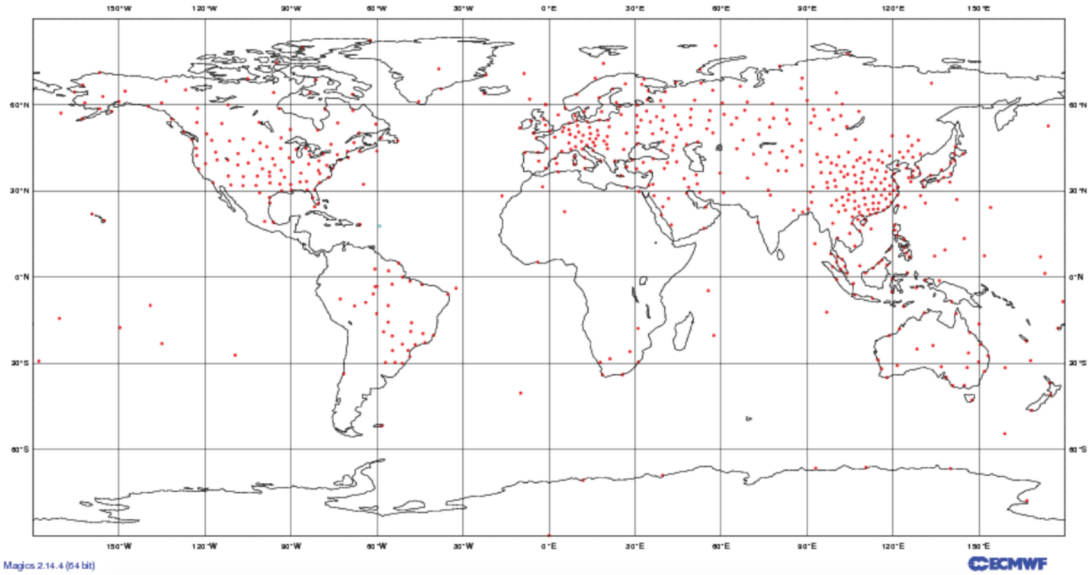



Figure 9.1.3: Upper air observation and sounding locations for temperature, pressure, and humidity aloft. Each point marks a station where radiosondes are released to measure atmospheric conditions, including temperature, humidity, pressure, and wind, at different heights. Launch sites are distributed across many countries on multiple continents, with higher concentrations in regions with established meteorological infrastructure and fewer sites in sparsely populated areas. The overall pattern illustrates the global coverage of the upper-air observing network used for weather forecasting and atmospheric analysis (CC-BY-NC-SA 4.0; Roland Stull, via Practical Meteorology).

## Sea-Level Pressure Adjustment

Weather stations exist at many altitudes. Because air pressure decreases with altitude, weather stations in cities at high elevations will report lower air pressure than those 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 sea-level pressure. If this were not done, pressure differences between nearby surface stations would be dominated by their elevation differences, and a surface pressure map would more closely resemble an elevation map than an atmospheric pressure map. By correcting for altitude differences, synoptic maps show mean sea-level pressure, which is used to depict low- and high-pressure centers near the surface.

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# 9.2: Synoptic Weather Maps

## The Station Model

Surface weather observations from many weather stations at a given time can be shown on a weather map using a **station plot model**, as shown in Figure 9.2.1. Station plot models typically display wind information via wind barbs, along with current weather conditions, sea-level pressure, temperature, dew point, visibility, and cloud cover. Sometimes, additional information, such as pressure tendency, cloud type, and precipitation amounts, is also provided.

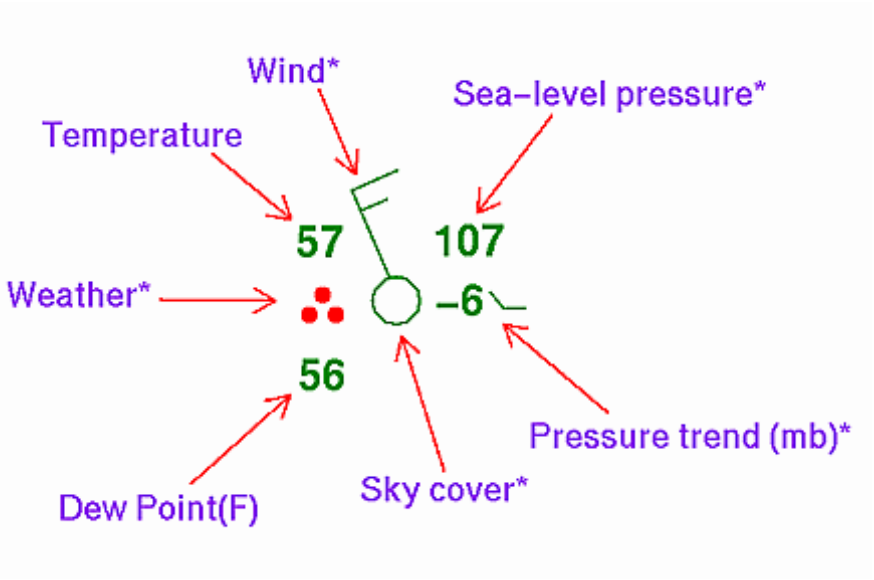



Figure 9.2.1: Diagram illustrating a standard meteorological station model used on weather maps. Labels identify each element of the station model and explain how multiple weather variables are encoded in a compact visual format for meteorological analysis. (Public Domain; The station plot via NOAA).

As illustrated in 9.2.1, the top left number 57 in the station model example indicates the current air temperature at the station's location in Fahrenheit (F). Similarly, the bottom-left number 56 indicates the dew point, the top-right number 107 indicates the sea-level pressure, and the bottom-right number -6 indicates the pressure trend. Additional symbols and numbers indicate current weather conditions and recent pressure changes.

At sea level, average pressure typically ranges from 950 to 1050 mb (millibars). With a decimal point, this can require four or five digits to be displayed. Instead, the station model abbreviates the pressure to three decimal places, with the last digit indicating the nearest tenth. The initial 9 or 10 is also left out. Thus, when a reading station models pressure, first place a decimal point to the left of the third (last) digit. If the three-digit number is less than 500, then a 10 is placed to the left of the number. If the three-digit number is greater than 500, then a 9 is placed to the left. So, “147” is actually 1014.7 mb, and “998” would be 999.8 mb.

Just below the sea-level pressure value is the pressure trend, measured in mb. It has two components: a number and a symbol to indicate how much (in tenths of millibars) the pressure changed over the past 3 hours and the trend in that change. In 9.2.1 , the pressure decreased by 6 mb and then became steady at the current value. Table 9.2.1 shows the meanings of the different pressure trend symbols.

*Table 9.2.1: Pressure trend symbols and their meanings.*

| Symbol | Meaning |
|---|---|
|  | Steady |
|  | Continuously Rising |
|  | Continuously Falling |
|  | Rising, then steady |
|  | Falling, then steady |
|  | Rising after slightly falling |
|  | Rising, then slowly falling |
|  | Falling after slightly rising |
|  | Falling, then slowly rising |

A barbed line extending from the station model's central circle represents wind, with its orientation and markings indicating wind direction and speed. 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. The barbs on the pole indicate wind speed. Table 9.2.2 illustrates what each wind barb symbol represents. The total wind speed is calculated by adding all the barbs and pennants. On many weather maps, wind speeds are given in knots, though different maps may use different units, so it is always good to check.

*Table 9.2.2: Wind barb table illustrating the symbols used to denote wind speeds.*

| Symbol | Wind speed | Description |
|---|---|---|
|  | Calm | two concentric circles |
|  | 1 - 2 knots | a line with no barbs |
|  | 5 knots | a half barb |
|  | 10 knots | a full barb |
|  | 50 knots | a pennant (triangle) |

The center symbol indicates cloud cover, and Table 9.2.3 links numerical sky-cover values measured in Oktas (eighths) to standardized sky-cover categories, abbreviations, and equivalent tenths of sky coverage. Categories range from clear sky to overcast conditions, through few clouds, scattered clouds, and broken clouds. An additional row describes cases in which the sky is obscured or not measured. Each category is paired with its corresponding circular symbol used on weather maps, enabling users to interpret sky-cover information from plotted station observations.

*Table 9.2.3: Sky cover symbols used in meteorological station models.*

| Sky cover scale in Oktas | Symbols | Name | Abbreviation | Sky cover fraction (tenths) |
|---|---|---|---|---|
| 0 |  | Sky clear | SKC | 0 |
| 1 |  | Few clouds | FEW | 1 |
| 2 |  | | | 2 to 3 |
| 3 |  | Scattered clouds | SCT | 4 |
| 4 |  | | | 5 |
| 5 |  | Broken clouds | BKN | 6 |

| Sky cover scale in Oktas | Symbols | Name | Abbreviation | Sky cover fraction (tenths) |
|---|---|---|---|---|
| 6 |  | | | 7 to 8 |
| 7 |  | | | 9 |
| 8 |  | Overcast | OVC | 10 |
| (9) |  | Sky obscured | | unknown |

Finally, in Figure 9.2.1, notice that the two temperatures on the left side are separated by a series of dots representing precipitation. An excellent guide to precipitation symbols can be found in Figure 9.2.2 and Table 9.2.4.




Figure 9.2.2: The NOAA convention showing precipitation type and intensity on a station model. (Public Domain; Station Model via NOAA/NWS)

*Table 9.2.4: Precipitation symbols on a station model and their meaning.*

| Symbols | Meaning |
|---|---|
| One, two, or three solid dots in a column | Rain (light, moderate, heavy) |
| One, two, or three asterisks in a column | Snow (light, moderate, heavy) |
| A lightning bolt with a dot or asterisk | Thunderstorm (with rain, snow, or no precipitation) |
| An inverted triangle with a dot or asterisk | Shower (rain or snow) |
| Two short apostrophe-like marks | Drizzle |
| Curved lines resembling the number 2 | Freezing rain or freezing drizzle |
| Triangle with a dot | Ice pellets or sleet |
| Two or three stacked horizontal bars | Fog (thin or thick) |

| Symbols | Meaning |
| --- | --- |
| Infinity symbol | Haze |

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# 9.3: Map Analysis

## Analyzing Weather Maps

Surface observations, as described above, are plotted on a map as station plot models, as shown in Figure 9.3.1. The raw data provides the prevailing wind direction in different parts of the US, which areas are experiencing widespread rain, and which areas are overcast. From the wind direction patterns, you can even infer centers of high and low pressure. However, much of this information is unclear without some analysis of the map. Map analysis can be done by hand or by computer and involves drawing **contours (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 atmospheric boundaries, such as **fronts** or **dry lines,** to show the locations and movement of air masses. Here, we will focus on lines of constant pressure (isobars) and lines of constant temperature (isotherms).

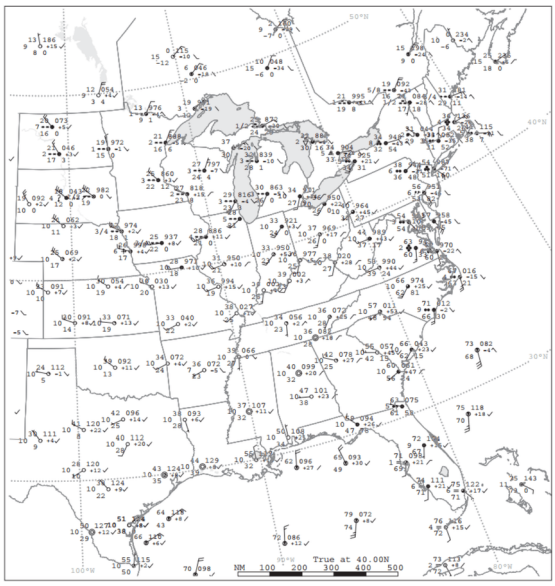


Figure 9.3.1: A surface weather map with numerous meteorological station models, each representing observations from an individual surface weather station. At each station, numerical values and symbols indicate air temperature, dew point, sea-level pressure, wind speed and direction, cloud cover, present weather, and recent pressure change. The dense distribution of station plots illustrates detailed, real-time surface weather conditions across the region, supporting weather monitoring, forecasting, and synoptic analysis. (CC-BY-NC-SA 4.0; Roland Stull via Practical Meteorology) .

## Isopleths

Figure 9.3.2 shows a simplified example of a surface temperature map. It shows a uniform grid of equally spaced temperature values in °C. Values range from the low single digits at one edge of the grid to the high twenties and low thirties at the opposite edge. The top row contains values between 2 and 5 degrees, while successive rows show progressively higher temperatures. Mid-grid values are mostly between 20 and 30 degrees, and the highest values, around 29 to 33 degrees, appear clustered near one side of the grid. When analyzing temperature on such a map, you draw isopleths of constant temperature, called isotherms. The main purpose is to separate areas of warmer air from those of cooler air and to determine how quickly temperature changes across a horizontal distance, 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.

Observing the temperature field in Figure 9.3.2, you can see that many values do not align 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 a value based on the data around it. Starting in the upper-left corner, imagine drawing your 5°C isotherm. Five lies much closer to 4 than 16, so you will start your isotherm close to 4. 5°C will also lie between 3 and 15°C, but slightly farther away from the 3 value, so you will slope your isotherm down slightly. 5 °C will also lie 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, slope back upward slightly between 3 and 6, cross directly through the middle between 4 and 6, and then back up through 5 in the upper right-hand corner.

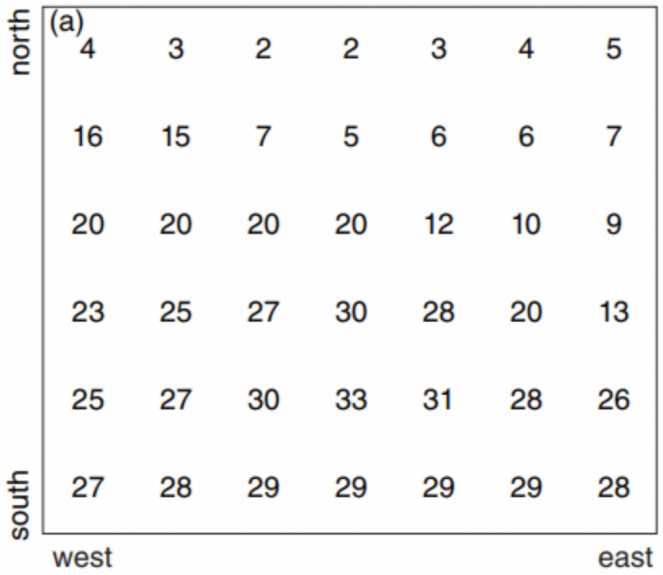



Figure 9.3.2: A simplified surface temperature map, with values in degrees Celsius arranged in evenly spaced rows and columns. The X and Y axes represent east-west and north-south extents on a map. The temperature values show a smooth increase across the region, indicating a spatial temperature gradient. (CC-BY-NC-SA 4.0; Roland Stull, via Practical Meteorology).

The complete contoured example of this temperature field is in Figure 9.3.3. 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 connect) or extend to the edge of the map. They will *never* cross or end suddenly. If two contours crossed, it would mean that at some point, one air temperature would differ from another, which is impossible. Areas where isotherms are close together indicate a strong temperature gradient, which may indicate a frontal zone.

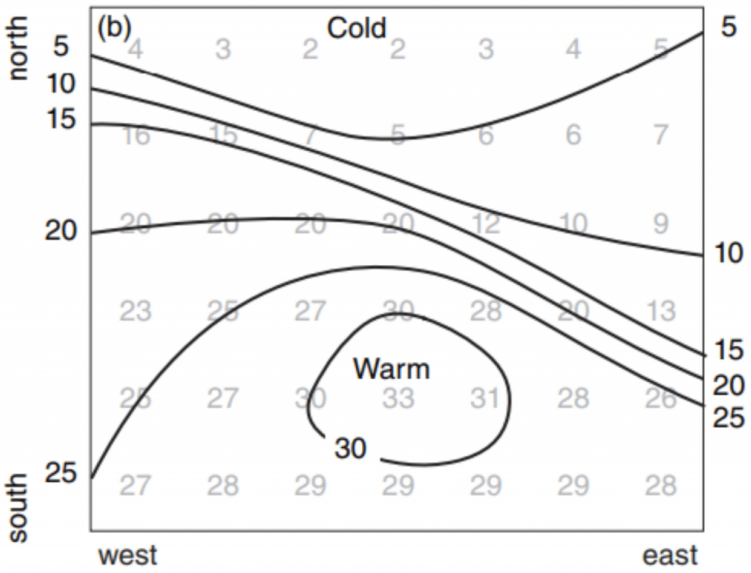



Figure 9.3.3: The smooth contour lines connect locations with equal temperature, illustrating gradual temperature changes across the region. A labeled cool region corresponds to lower-temperature values, grouped by higher-frequency isotherms, while a labeled warm region encloses the highest temperatures, including a closed contour around values near 30 degrees and above. (CC-BY-NC-SA 4.0; Roland Stull, via Practical Meteorology).

## 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 to identify areas of high and low pressure, as shown in Figure 9.3.4. Areas where isobars are drawn close together indicate strong pressure gradient forces. A stronger pressure gradient indicates higher 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.

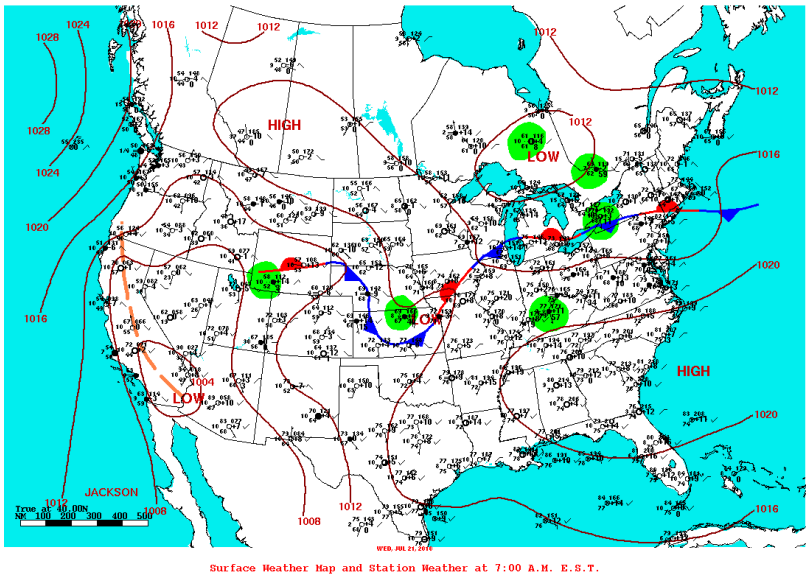



Figure 9.3.4: Surface weather map of continental North America at 7 am (EST) on July 21, 2010. A low-pressure trough spans the length of California, while high-pressure regions exist off the coast of South Carolina and north of Montana. (Public Domain, National Center for Environmental Prediction via Wikimedia Commons).

Figure 9.3.4 displays numerous surface weather station models, each encoding observed conditions such as air temperature, dew point, wind, cloud cover, sea-level pressure, and present weather. Curved contour lines represent lines of equal sea-level pressure,

illustrating the overall pressure pattern across the region. Labeled pressure systems identify areas of relatively higher pressure and lower pressure, each associated with surrounding isobars. Frontal boundaries are drawn using standard meteorological symbols to indicate cold, warm, and stationary fronts, highlighting zones of air-mass interaction. The combined features depict a snapshot of atmospheric conditions used for synoptic-scale weather analysis and forecasting at a specific observation time.

Because weather information is shared across continents and country borders, it is important to maintain standards so that everyone 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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# 9.4: Exercises for Chapter 9

## Chapter 9: Exercises

1. Decode the following METARs, giving Air Temperature (in °C), Dew Point Temperature (in °C), Air Pressure (in Inches of Mercury), Cloud Cover (both amount and height), Wind Direction (in degrees), and Wind Speed (feel free to Google search the IACO airport code for each airport if you are interested in seeing what airport you are decoding a METAR for):
   a. KSJC 042053Z 14009KT 10SM FEW008 BKN028 OVC038 12/11 A2987 RMK AO2 RAE52 SLP113 P0003 60036 T01220106 58011
   b. PHNL 052053Z 07008KT 10SM FEW024 SCT040 BKN065 26/20 A3002 RMK AO2 SLP166 T02610200 50001
   c. KDEN 070153Z 26013KT 10SM CLR 08/M14 A2987 RMK AO2 SLP094 T00781139
2. Consider the map below, Figure 9.4.1. Decode the station models for cities a, b, c, d, and e, giving Air Temperature (in °F), Dew Point Temperature (in °F), Air Pressure (in Millibars), Cloud Cover (amount), Wind Direction (cardinal direction), and Wind Speed (in kts). NOTE: City b. has a wind speed indicated by one long feather and one short feather while city c. has a wind speed indicated by one short feather:

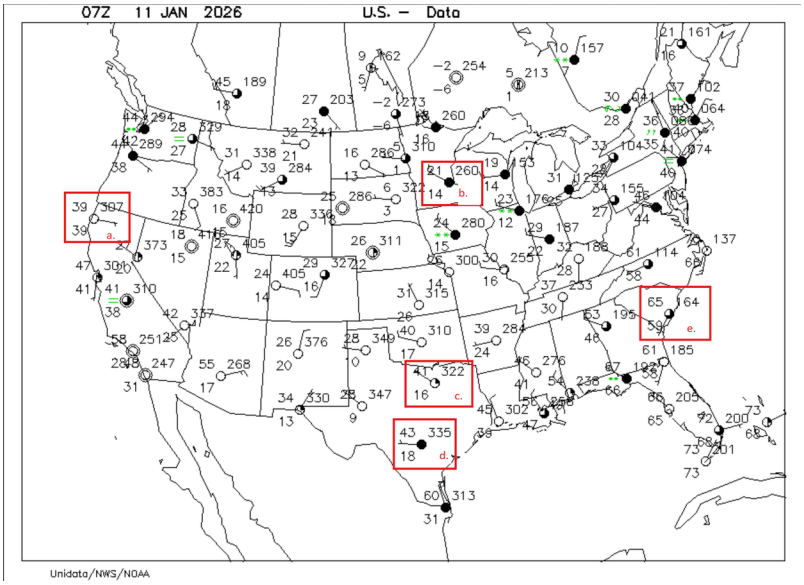



Figure 9.4.1: Surface weather observations. Alternative description in the text below (CC BY-NC 4.0; American Meteorological Society via Unidata).

Figure 9.4.1 is a Surface weather observation map of the contiguous United States showing station models plotted at reporting locations for a labeled observation time and date. State boundaries and coastlines provide geographic context. Each station model includes a central circle representing cloud cover, a wind barb showing surface wind, and surrounding numbers encoding surface variables such as air temperature, dew point, and sea-level pressure. Several individual stations are enclosed by rectangular callouts labeled with letters, indicating selected examples for closer interpretation. These highlighted stations demonstrate how surface weather information is encoded at specific locations. The map provides a snapshot of surface atmospheric conditions across the country for instructional analysis and forecasting.

**Instructor's Note:** Exercise 2 may not be fully accessible for all students with visual impairments, so it is suggestion that you have an alternative exercise prepared in case you have a visually impaired student who is unable to complete Exercise 2.

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# 10: THUNDERSTORMS

## Introduction

A thunderstorm is a rainstorm accompanied by thunder and lightning. Figure 10.1 depicts lightning and dark clouds during a thunderstorm in northern Sicily. Thunderstorms are deep convective clouds that have a significant vertical extent all the way from the boundary layer to the tropopause. Thunderstorms often bring severe weather, including heavy rain, hail, lightning, damaging winds, and occasionally tornadoes. Thunderstorms are sometimes thought to be synonymous with cumulonimbus clouds, though not all cumulonimbus clouds are thunderstorms.



Figure 10.1: Lightning strikes the water from a thunderstorm in Northern Sicily. (CC BY-SA 2.0; Gerlos via Wikimedia Commons).

The primary energy driving thunderstorms is the conversion of warm, moist air into clouds and precipitation, which releases significant latent heat during condensation. 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 help warm air parcels gain the initial buoyancy needed to begin rising. Thunderstorms can last anywhere from less than an hour to more than 12 hours. Due to their rapid growth, relatively small size (mesoscale), and sensitivity to environmental conditions, it is challenging to predict a storm's exact time and location. 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 precisely where or when it will occur.

### Learning Objectives

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

- Name and label parts of a thunderstorm.
- Describe the importance of updraft, downdraft, and windshear for a thunderstorm.
- Identify a few types of thunderstorms (airmass, MCC, squall line, supercell, etc.).
- Discuss a few favorable conditions for thunderstorm formation.

10.1: Thunderstorms formation and types

10.2: Precipitation during Thunderstorms

10.3: Winds during thunderstorms
10.4: Lightning and Thunder
10.5: Tornadoes
10.6: Keeping Yourself Safe
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# 10.1: Thunderstorms formation and types

## Thunderstorm Life Cycle

The three main stages of a thunderstorm are illustrated in Figure 10.1.1, and are as follows:

1. Developing (cumulus) stage
2. Mature stage
3. Dissipating stage

In the **developing stage**, rising air currents, known as "**updrafts,**" are dominant, and cumulus clouds with some vertical growth are formed. This type of cloud is also known as a cumulus congestus. Next is the **mature stage,** where the growth continues, resulting in a towering cumulus, known as 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, creating a descending air current within 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 is accompanied by thunder and lightning, it is referred to as a **thunderstorm**. A charge separation in clouds causes lightning. For this charge separation to occur, a cloud must have ice crystals inside of it, which is why thunderstorms are always mixed-phase or contain both ice and liquid droplets.

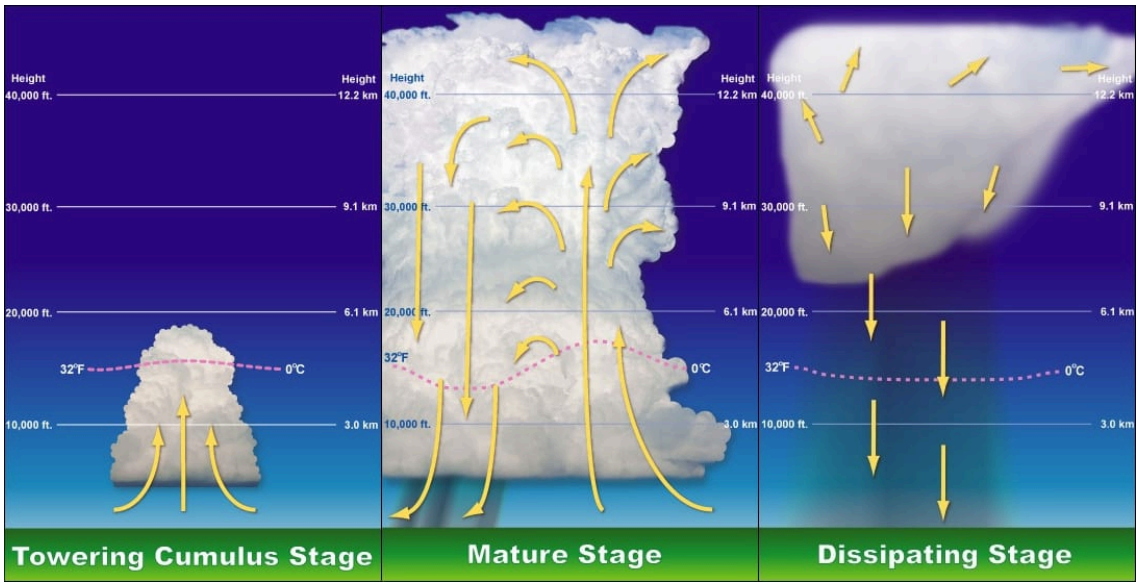



Figure 10.1.1: The three stages of thunderstorm formation include the developing, mature, and dissipating stages. (Public Domain; NOAA via Wikimedia Commons). Alternative Description of Image.

## Airmass Thunderstorm

The thunderstorm described in Figure 10.1.1 is called an airmass thunderstorm. In general, the ingredients necessary for a thunderstorm to occur are **warm, moist air**, some **instability**, and a **trigger** that can initiate the storm system. Airmass thunderstorms typically occur in environments with similar properties and little wind shear. They typically occur in summer, are benign, and last for only a few hours. They are also known as **single-cell thunderstorms**, where one updraft and one downdraft together form a convective cell. The reason airmass thunderstorms are so short-lived and benign is that they lose the ability to maintain themselves. This is due to two primary reasons:

1. Below the 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 that drives further convection, the storm loses strength

and dissipates.
2. Rainfall within the storm updraft reduces the storm's strength and further dissipates it.

Another diagram of an air mass thunderstorm or single cell thunderstorm is illustrated in Figure 10.1.2. The initial small cumulus cloud develops into a cumulus congestus, characterized by predominantly updrafts. It then develops further into a mature cumulonimbus cloud, characterized by well-defined updrafts and downdrafts. An **overshooting top** often occurs at the top of the thunderstorm during the mature stage. Overshooting tops are dome-shaped due to strong updrafts in intense thunderstorms that reach the tropopause and then push 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. The thunderstorm is most intense during its mature stage, characterized by heavy rainfall, lightning, and thunder. The final stage is the dissipating stage, characterized by predominantly downdrafts. During the dissipating stage, the towering cumulonimbus begins to evaporate from the base upward, as downdrafts cool the surface and deprive the storm of its heat source. Often, only the anvil top remains after dissipation.

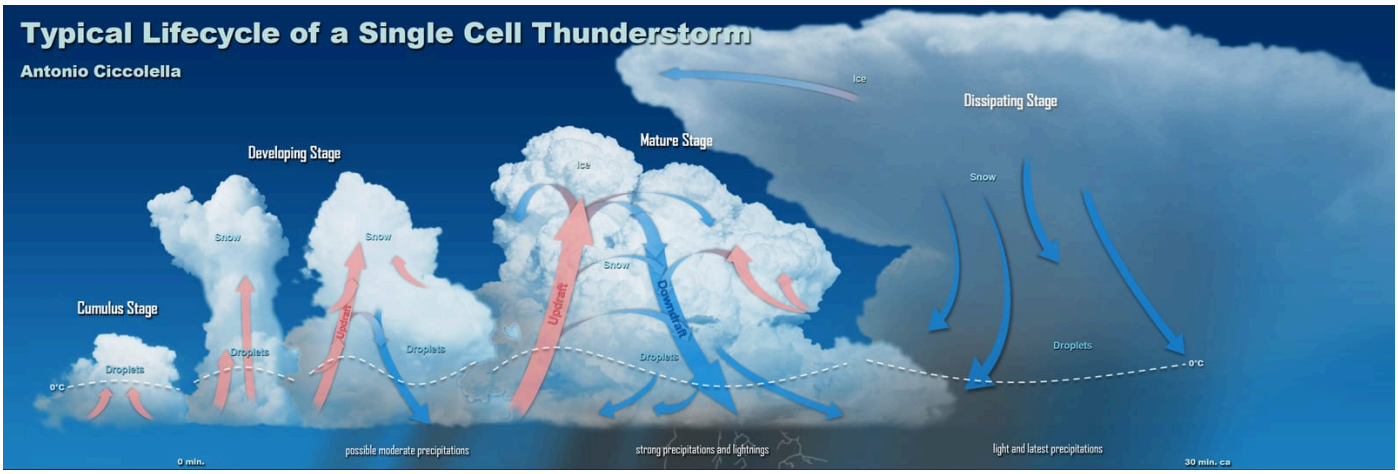



Figure 10.1.2: The typical life cycle of a thunderstorm consists of primarily updrafts during the developing stage, both updrafts and downdrafts during the mature stage, and primarily downdrafts during the dissipating stage. (CC BY-SA 4.0; Antonio Ciccolella via Wikimedia Commons).
Alternative description of the image

The trigger for a thunderstorm can be a warm, humid air mass heated from below by daytime solar heating or forced to rise by terrain (orographic thunderstorms). The collision of air masses along a warm front, a cold front, or a dryline can also be a trigger.

## Severe Thunderstorm

While basic airmass thunderstorms typically consist of one “cell”, severe storms often contain multiple cells, each with a separate rising thermal associated with the same storm system. In a multicell storm, each cell typically occupies a different life-cycle stage and continues to develop through successive stages.

For a thunderstorm to become severe, one additional important ingredient is necessary. In addition to warm, moist air, some 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 that it helps 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 longer than air-mass thunderstorms. Severe thunderstorms take many forms.

## 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 line of linearly propagating severe thunderstorms. Squall lines consist of a long, narrow line of merged thunderstorms, often triggered by a cold front, dry line, or gust front as depicted in Figure 10.1.3. Squall lines can be many hundreds of kilometers long, but their width is usually much shorter, ranging from 15 to 100 km. They can sustain themselves for

several hours to several days.

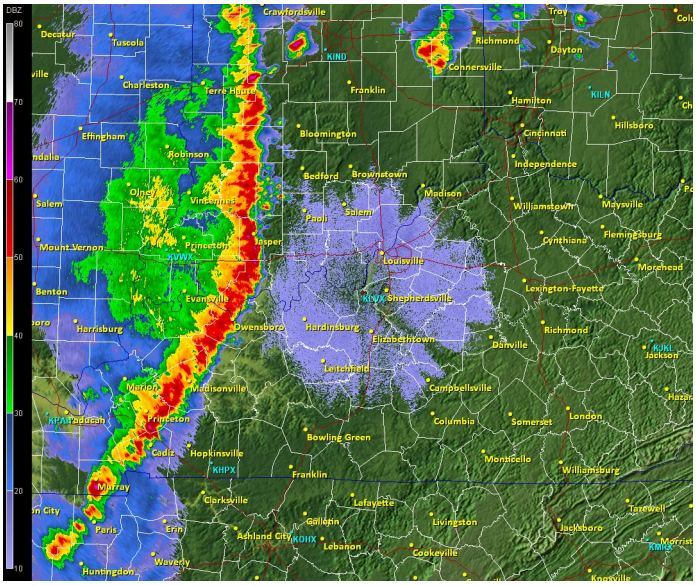



Figure 10.1.3: A squall line is a line of thunderstorms, often forming along or ahead of a cold front. (Public Domain; Severe Weather 101 via NOAA/NSSL). Alternative description of the image.

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 Earth's surface, it spreads laterally, much like 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 Systems

A Mesoscale Convective System (MCS) is a collection of thunderstorms that act together as a system, as depicted in Figure 10.1.4. An MCS can span an entire state and persist for more than 12 hours. On radar, this system may appear as a solid line, a broken line, or a cluster of cells.

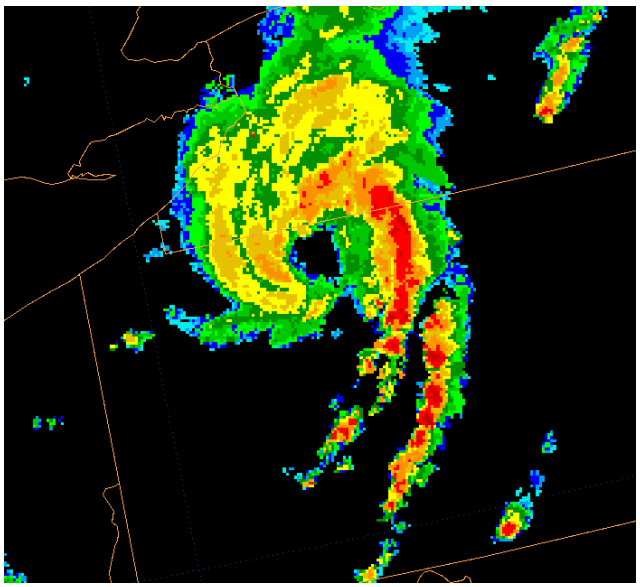


Figure 10.1.4: Radar reflectivity from a Mesoscale Convective System over Pennsylvania with a trailing squall line. (Public Domain; David Roth via Wikimedia Commons). Alternative description of the image.

**A mesoscale convective complex (MCC)** is a collection of thunderstorms of the severe type. The storm system features a cloud shield (anvil) with a diameter of at least 350 km, typically elliptical or circular, and lasts 6-12 hours. MCCs are huge storms that occur multiple times per year, especially in the central United States. They are truly impressive convective masses.

## Supercell

A **supercell** is a type of severe storm characterized by rotation as illustrated in Figure 10.1.5. 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 clearly defined structure that enables them to produce tornadoes. The precipitation is located below the shelf cloud and away from the tornado. Supercells have additional features, such as 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. Just like other cumulonimbus clouds, the supercell has an anvil near the top. But the updraft in such a rotating mega storm can be strong enough to overshoot the anvil and deposit moisture into the stratosphere.

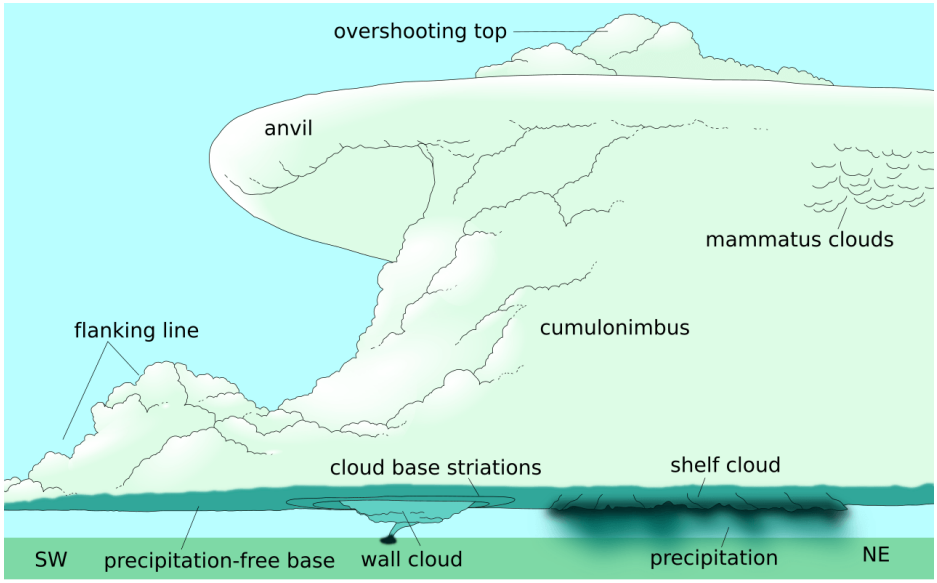



Figure 10.1.5: A Supercell consists of a rotating cumulonimbus core with precipitation, such as light rain, heavy rain, and hail, pushed downwind by wind shear. The top of the cloud is an anvil-shaped feature, with some overshooting top clouds. (CC BY-SA 3.0; Vanessa Ezekowitz via Wikimedia Commons). Alternative description of the image.

Figure 10.1.6 is an annotated version of an actual supercell thunderstorm that created a tornado. Remember that supercells are rotating storm systems that can produce tornadoes, but they don't always. In fact, very few supercells produce tornadoes, but the ones that do receive the most attention, since they are the most destructive.

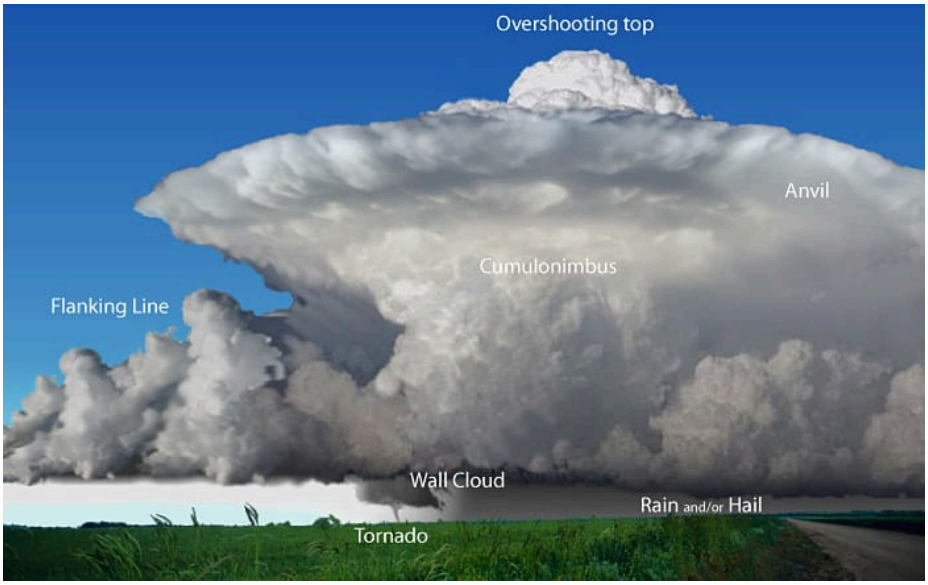



Figure 10.1.6: An annotated image of a supercell cumulonimbus cloud depicting the wall cloud and tornado near the base and an anvil near the top. Precipitation occurs away from the rotating core. (Public Domain; NOAA via Wikimedia Commons). Alternative description of the image.

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# 10.2: Precipitation during Thunderstorms

## Types of precipitation during thunderstorms

Thunderstorms consist of deep convective clouds that can produce large raindrops (2–8 mm in diameter) within 5–10 km wide rain shafts that move horizontally across the ground, with precipitation lasting from 1 to 20 minutes over a fixed point. Rainfall rates may be high, with rates ranging from 10 to 1000 mm per hour. Figure 10.2.1 depicts heavy rain shafts produced by a supercell thunderstorm in Montana. However, thunderstorms can generate a variety of precipitation types.



Figure 10.2.1: A supercell thunderstorm in Montana producing heavy, localized precipitation. (CC BY 2.0; Jedimentat44 via Flickr).

### Graupel

The cloud tops of thunderstorms extend far up in the troposphere. In the upper regions of the clouds, ice-phase processes take place. In the cloud, ice crystals and water droplets of different sizes exist, and heavier hydrometeors fall faster than smaller ones, sometimes colliding and coalescing. If these heavier ice particles fall through regions of supercooled droplets (liquid water droplets below freezing), they may grow through a process called **riming**. Liquid water droplets freeze instantly and adhere to the outer surface of falling ice particles upon contact. This forms snow pellets or soft hail called **graupel**, which are less than 5 mm in diameter. Fallen graupel is pictured in Figure 10.2.2.



Figure 10.2.2: Graupel pellets on a car roof from the previous evening's thunderstorm in Elko, Nevada. (CC BY-SA 3.0; Famartin via Wikimedia Commons).

Alternatively, smaller ice crystals that fall just below 0°C may partially melt and adhere to other partially melted ice crystals via collisions, thereby promoting their growth through **aggregation**. These snow aggregates can reach up to 1 cm in diameter. Snow aggregates and graupel can sometimes reach the ground still partially frozen, even in summer, if they fall within cooler, saturated downdrafts. More frequently, these larger ice particles will melt completely into large raindrops before hitting the ground.

## Hail

**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 can carry smaller ice particles far above the freezing level, where collisions with supercooled droplets enable 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 regions of the cloud with high liquid water content, additional ice coatings 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 a hailstone is carried away from the updraft or becomes too heavy, it will fall because the rising air no longer supports it.

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. Figure 10.2.3 shows people holding some hailstones which are cut in half. Note in the figure that the hailstone has concentric layers, similar to those found in tree rings. The concentric layers alternate between clear and opaque ice. This relates to the hailstone's path through the cloud and the liquid water content of the regions it passes through. In areas with low liquid water content (a dry growth regime), supercooled water droplets freeze instantly onto the hailstone, producing a milky, opaque ice that contains many air bubbles. In areas with high liquid-water content (wet-growth regime), supercooled water droplets accumulate rapidly on the hailstone. The transition from liquid to ice releases latent heat, thereby maintaining the hailstone's surface temperature at 0°C. As a result, supercooled droplets will form a water film around the stone rather than freezing immediately. 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.



Figure 10.2.3; Golf ball-sized hail cut in half so the inside concentric layers can be viewed. (Public Domain; NOAA via Wikimedia Commons).

Large hail can cause severe damage to crops, vegetation, aircraft, roofs, and windows, and create a traffic hazard if hailstones accumulate on roadways. 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 m/s (112 mph).

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# 10.3: Winds during thunderstorms

## Gust Fronts

Recall that mature thunderstorms contain both updrafts of warm, buoyant moist air and 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 in all directions along the surface, as illustrated in Figure 10.3.1. This surface boundary between the cold, horizontally rushing air and the warmer air around it is called the **gust front**. Along the base of the cloud, there oftentimes exists a rotating section of cloud along the boundary between the warm inflow and cold downdraft called the **roll cloud**.

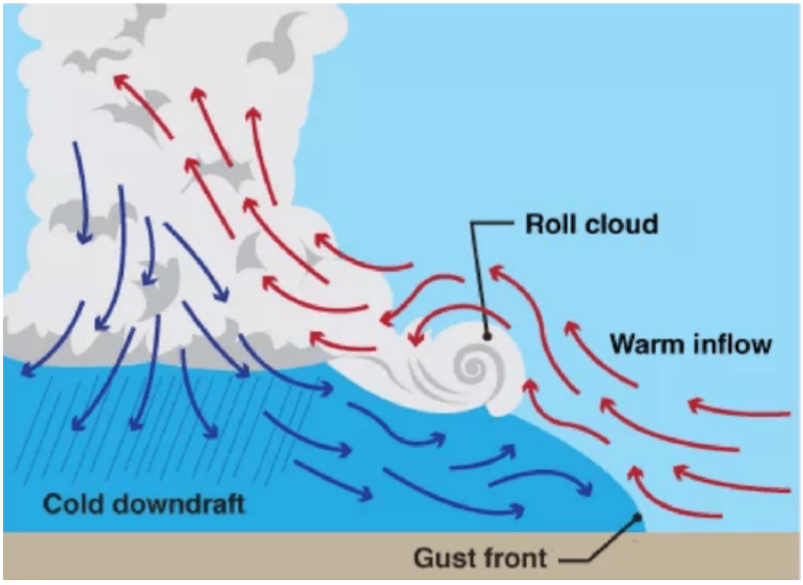



Figure 10.3.1: Schematic of a thunderstorm base showing a gust front formed by horizontally spreading downdrafts. The gust front separates the cold downdraft from the thunderstorm and the warm convective flow. A roll cloud may form along the gust front due to the strong wind shear. (Public Domain; What is a Gust Front via NOAA)

Along this boundary, winds change rapidly in both 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 Figure 10.3.2.



Figure 10.3.2: A shelf, or arcus cloud, appearing as a wedge-shaped cloud formed on the edge of a thunderstorm gust front over a city. (CC BY-SA 3.0; John Kerstholt via Wikimedia Commons).

When the atmosphere is conditionally unstable, the gust front may produce additional multicell thunderstorms along its leading edge by forcing warm, moist air upward. Each of these additional thunderstorms will have its own gust front, which may merge with other gust fronts to form a large gust front known as an **outflow boundary**.

## Downbursts and Microbursts

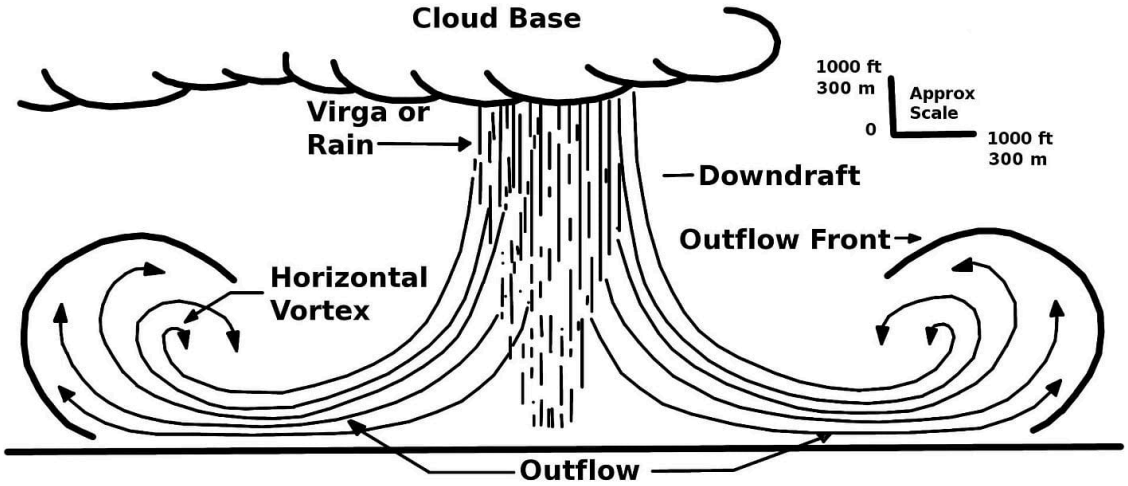



Figure 10.3.3: Cross-section of a microburst showing a rapid outflow caused by a strong downdraft of cold air from the cloud base. The outflow spreads horizontally away from the precipitation, producing a gust front or outflow front. Horizontal vortexes may form as the outflow front curves upward due to interaction with the rising warm air near the surface. (Public Domain; USFAA via Wikimedia Commons).

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 windburst, analogous to water rushing from a faucet and striking the bottom of the sink. This sort of downdraft is called a **downburst**. A short downburst, with width of 4 km (approximately 2.5 miles) or less, is called a **microburst,** as illustrated in Figure 10.3.3. A microburst consists of a downdraft of rain (or virga) falling out of the cloud base, which flows outward when it reaches the ground. As the outflow extends horizontally, horizontal vortexes develop on the edges of the outflow, known as the outflow front. Despite their small spatial extent, strong microbursts can generate winds exceeding 146 knots, capable of downing trees and severely damaging 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 can also occur with ordinary cell thunderstorms and even with clouds that produce only scattered showers.

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# 10.4: Lightning and Thunder

## Lightning

**Lightning** is caused by a discharge of electricity due to charge separation that typically occurs in mature thunderstorms. The main charging area in a thunderstorm occurs in the central part of the storm where air is moving upward rapidly (updraft) and temperatures range from -15 to -25 Celsius, as illustrated in Figure 10.4.1.

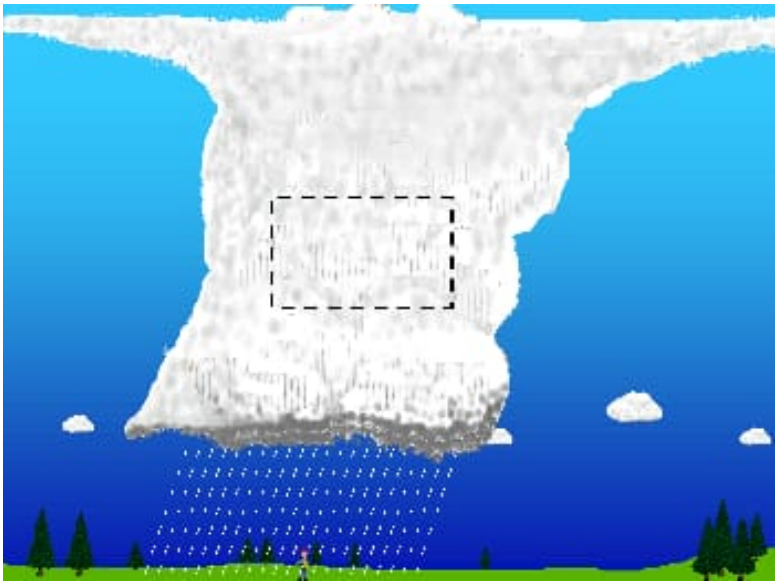


Figure 10.4.1: Strong updrafts brush against precipitating particles and result in charge separation in the centre of the cumulonimbus tower. (Public Domain; Understanding Lightning via NOAA/NWS)

At that place, the combination of temperature and rapid upward air movement produces a mixture of supercooled cloud droplets (small water droplets below freezing), small ice crystals, and soft hail (graupel). The updraft transports supercooled cloud droplets and very small ice crystals upward. At the same time, the graupel, which is considerably larger and denser, tends to fall or be suspended in the rising air. Differences in precipitation movement cause collisions. When the rising ice crystals collide with graupel, the ice crystals become positively charged, and the graupel becomes negatively charged.

The updraft carries the positively charged ice crystals upward toward the top of the storm cloud. The larger and denser graupel is either suspended within the thunderstorm cloud or falls toward the lower part of the storm. The result is that the upper part of the thunderstorm cloud becomes positively charged while the middle to lower part of the thunderstorm cloud becomes negatively charged. While this is the main charging process for the thunderstorm cloud, some of these charges can be redistributed by air movements within the storm (updrafts and downdrafts).

The buildup of charge influences the charge on the ground beneath the clouds. Normally, the ground has a slight negative charge. However, when a thunderstorm is directly overhead, the large negative charge near the base of the storm cloud repels negatively charged particles on the ground beneath it. This causes the ground and any objects (or people) directly underneath the storm to become positively charged, as depicted in Figure 10.4.2.



Figure 10.4.2: People's hair often stands on end during thunderstorms because positive charge enters our body from the ground. The positive charge on the hair is attracted to the negative charges in the cloud. (Public Domain; Understanding Lightning via NOAA/NWS)

As the negative charge in the cloud increases, the ground becomes more positively charged. As the charge difference increases, positively charged particles are repelled by taller objects such as trees, houses, and telephone poles. A channel of negative charge, called a "stepped leader", will descend from the bottom of the storm toward the ground. It is invisible to the human eye and shoots to the ground in a series of rapid steps, each occurring in less time than it takes to blink your eye. As the negative leader approaches the ground, a positive charge accumulates on the ground and on objects on it.

This positive charge "reaches" toward the approaching negative charge via its own channel, called a "streamer". When these channels connect, the resulting electrical transfer is what we see as lightning. These steps are illustrated in Figure 10.4.3. After the initial lightning strike, if enough charge is left over, additional lightning strokes will use the same channel and will give the bolt its flickering appearance.

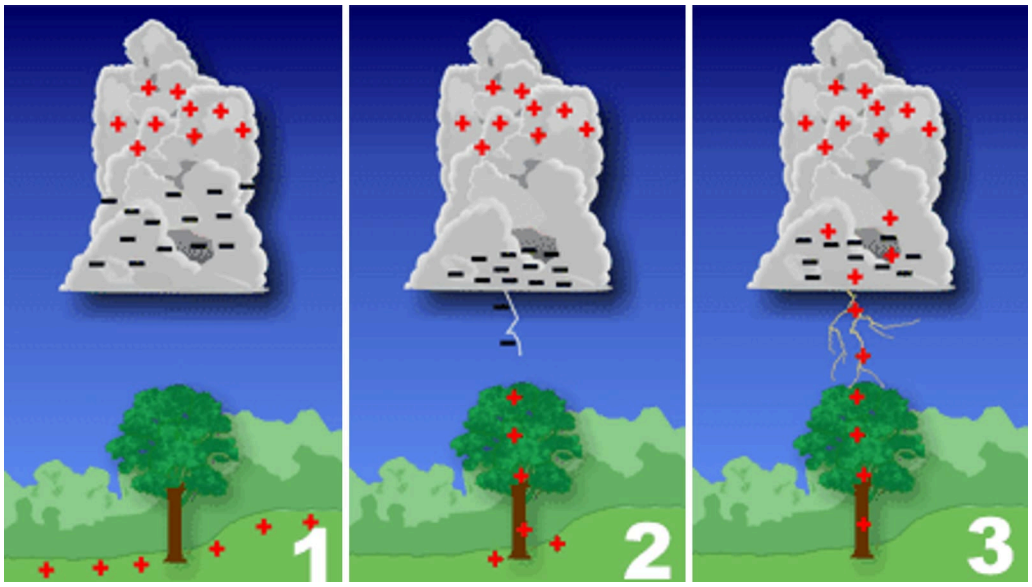



Figure 10.4.3: The three-step process of a cloud-to-ground lightning strike. In step 1, a positive charge accumulates on the ground near a negatively charged base of a thunderstorm. In step 2, a stepped leader descends from the bottom of the storm toward the ground, while the positive charge from the ground reaches out to meet the leader. When negatively and positively charged particles come into contact, a lightning strike occurs. (Public Domain; How lightning develops via NOAA/NWS)

Tall objects such as trees and skyscrapers are commonly struck by lightning. Mountains also make good targets. This is because their tops are closer to the base of the storm cloud. However, this does not always mean tall objects will be struck. It all depends

on where the charges accumulate. Lightning can strike the ground in an open field even if the tree line is close by. Lightning can also strike within the same cloud, from one cloud to another, or from a cloud to open air. Only about 20% of lightning strikes are cloud-to-ground strikes, while the majority of strikes occur within the cloud.

As a lightning strike travels through the atmosphere, it heats the surrounding air to a temperature of approximately 30,000°C, which is roughly five times hotter than the surface of the Sun. This intense heating causes a sudden, explosive expansion of the air, generating shock waves radiating in all directions from the flash. This expands outward as a sound wave called **thunder**. Due to the speed at which light travels, we see lightning instantly as it happens. Thunder travels much more slowly at about 330 $m \cdot s^{-1}$, 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. Sound travels 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.

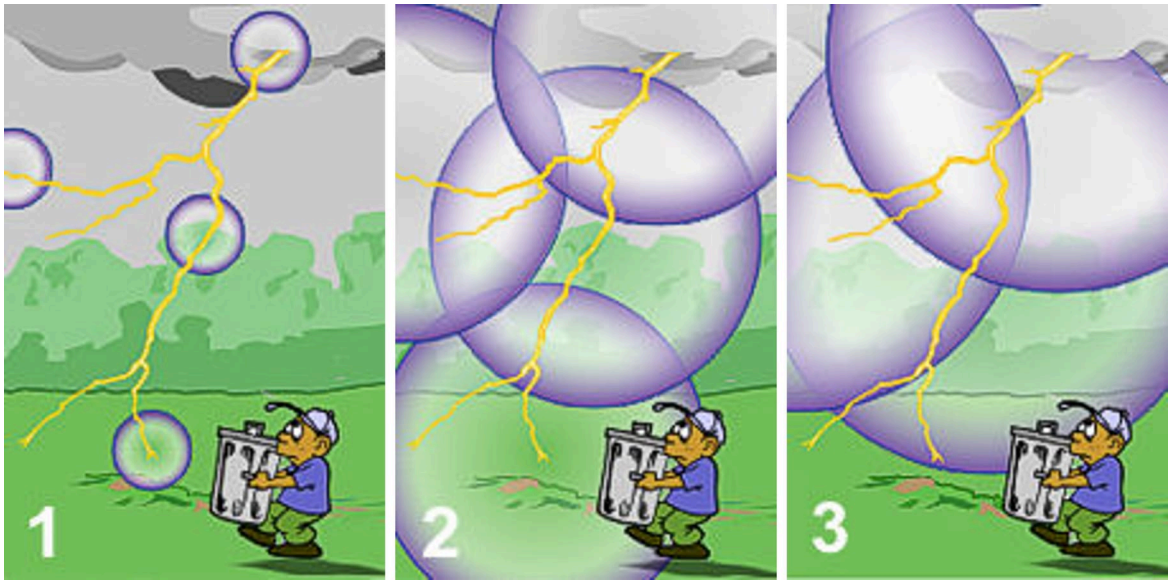



Figure 10.4.4: Rapidly expanding air within a lightning strike creates a shockwave in all directions. As the shockwaves from multiple locations along a lightning arc converge, the sound of thunder becomes a rumble. (Public Domain; The Sound of Thunder via NOAA/NWS)

When lightning strikes, a shock wave is generated at each point along the path of the lightning bolt. In the presence of nearby lightning, thunder will sound like a loud bang, crack, or snap, and its duration will be very short. As the shock wave propagates away from the strike centre, it stretches, attenuates, and elongates. Subsequently, shock waves from more distant locations reach the listener. To the listener, the combination of shock waves gives thunder its continuous rumble.

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# 10.5: Tornadoes

## What is a Tornado?

A tornado, also known as a twister, is a rapidly rotating column of air that extends vertically from the surface of the Earth to the base of a thunderstorm. 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 is driven by directional wind shear in the environment. Occasionally, a supercell storm produces a tornado due to rapidly rotating air in cyclostrophic balance. We learned in Chapter 8 that cyclostrophic balance occurs when the pressure-gradient force equals 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. Figure 10.5.1 depicts a tornado in Elie, Manitoba, in June 2007.



Figure 10.5.1: F5 tornado viewed from the southeast as it approached Elie, Manitoba, on Friday, June 22, 2007. (CC BY-SA 3.0; Justin1569 via Wikimedia Commons.)

## Predicting Tornadoes

Tornadoes are difficult to forecast with sufficient lead time. Typically, a Tornado Watch will be issued by the 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 observed on a weather radar, as illustrated in Figure 10.5.2 . The key feature is the bottom-left region of the storm, where a rain band wraps around a region of dry inflow air. At the cusp of the hook, a tornado would form if it were occurring. The strong rotation in a tornado creates this radar signature. 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.

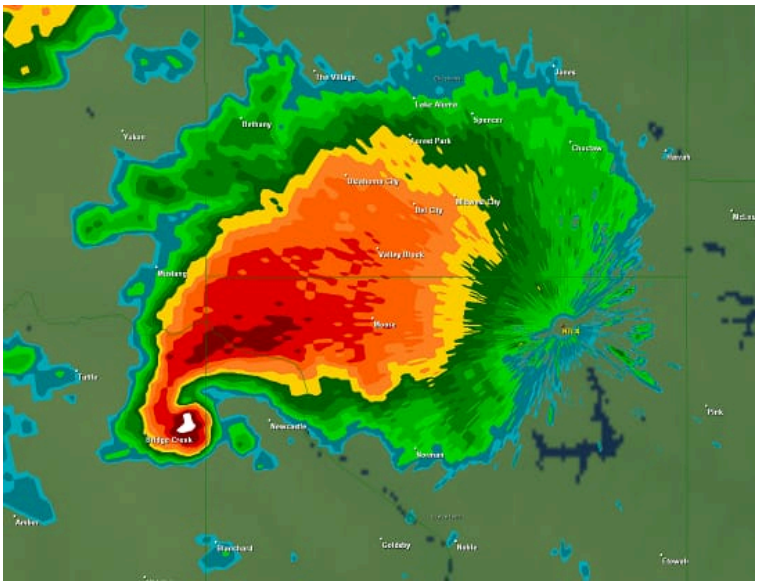


Figure 10.5.2: Hook echo appears on a weather radar when strong rotation in a tornado flings rain and ice in a spiral pattern around a dry air region. (Public Domain; White Fox via Wikimedia Commons).

## The Fujita Scale

The Fujita Scale is a scale for rating tornado intensity, based primarily on the damage tornadoes inflict on human-built structures and vegetation. Thus, the strength or scale of a Tornado is categorized only after it occurs. The scale was introduced in 1971 by Dr. Tetsuya Fujita of the University of Chicago, in collaboration with Allen Pearson, then head of the National Severe Storms Forecast Center (now the Storm Prediction Center). In 2007, the Fujita Scale was replaced by the **Enhanced Fujita Scale (EF),** a modified version of the original scale that is currently used to classify tornado strength and wind speed based on observed damage. Table 10.5.1 lists the six categories of the EF scale along with the associated wind speed and damage.

*Table 10.5.1: The 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 |

## Tornado Occurrence

Tornadoes occur in many regions worldwide, including Australia, Europe, Africa, Asia, and South America. Even New Zealand reports about 20 tornadoes each year. Two of the highest concentrations of tornadoes outside the U.S. are in Argentina and Bangladesh. But the central US is the world's tornado hot spot. Approximately 1,200 tornadoes occur in the U.S. each year. Because official tornado records date back only to 1950, we do not know the actual annual average number of tornadoes. Additionally, tornado spotting and reporting methods have undergone significant changes over the last several decades, resulting in a greater number of tornadoes being observed.

**Tornado Alley** refers to a broad area of relatively high tornado occurrence in the central United States. As a colloquial term, there are no definitively set boundaries, as illustrated in Figure 10.5.3. Still, the area common to most definitions extends from Texas through Oklahoma, Kansas, Nebraska, South Dakota, Iowa, Minnesota, Wisconsin, Illinois, Indiana, Missouri, Arkansas, North Dakota, Montana, Ohio, and eastern portions of Colorado, New Mexico, and Wyoming.

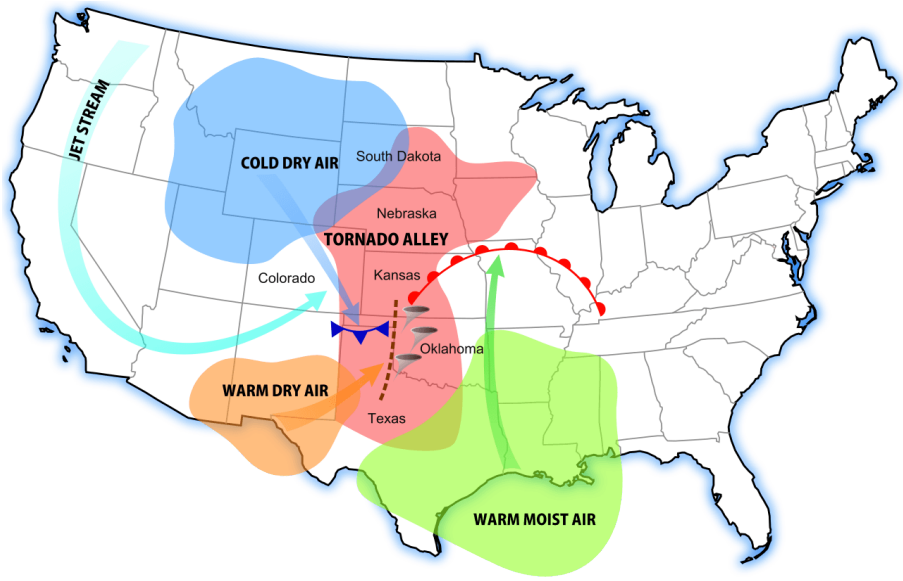



Figure 10.5.3: A unique interaction occurs when dry cold air moving south from Canada and northwestern U.S. meets warm moist air traveling north from the Gulf of Mexico. When the jet stream is positioned over this area (Specifically when the eastern half of an upper-level trough in the jet steam present), a mid-latitude cyclone develops. Severe thunderstorms form along a long trough extending from the center of the mid-latitude cylone over central Kansas down to central Texas. Thus, states like Texas, Oklahoma, Kansas, and Nebraska experience numerous tornadoes annually. This region is commonly known as Tornado Alley, which is sandwiched between the warm moist air from the Gulf of Mexico, and the cold dry air from Canada and the Northwestern U.S.. (CC BY-SA 3.0; Dan Craggs via Wikimedia Commons)

This region is particularly conducive, since it has all the necessary ingredients for tornado formation:

1. **Flat, dry terrain:** Tornadoes are more easily formed in regions with flat, dry terrain.
2. **Atmospheric instability:** Warm, moist air from the Gulf of Mexico combines with cool, dry Rocky Mountain winds.
3. **Competing air masses:** When competing air masses converge, warm air rises and cool air sinks, producing severe thunderstorms and sometimes supercells. If the two air masses clash at different speeds or in different directions, a funnel cloud can form, leading to a tornado.

While Tornado Alley is widely recognized as a tornado hotspot, tornadoes and tornado-like features can occur elsewhere. A dust devil and a waterspout are tornado-like features that are associated with cumuliform and sometimes cumulonimbus clouds. However, dust devils and waterspouts are typically not associated with supercells. Waterspouts look like tornadoes but form over water surfaces and do not reach the destructive wind speeds of tornadoes.

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# 10.6: Keeping Yourself Safe

## Safety during Thunderstorms

From reading this chapter, hopefully, you've learned that severe storms and storm hazards should be taken seriously. One of the challenges in severe weather forecasting is that the impacts are often extremely localized. It is difficult to know exactly where hail will fall or a tornado will pass. Often, you'll be aware that the possibility exists for storm impacts. Therefore, it is important to take precautions, regardless of whether you are in the right place at the wrong time.

## Lightning Safety

Lightning is one of the most dangerous aspects of a thunderstorm. Lightning can strike up to 10 miles from the main area of the thunderstorm. That is about the distance you can hear thunder from the storm. Whether or not you can see the actual lightning flash, if you can hear thunder, you are at risk of being struck by lightning. Because of this, the National Weather Service has adopted the following motto: ***"When Thunder Roars, Go Indoors ... and stay there at least 30 minutes after the last clap of thunder."*** Figure 10.6.1 illustrates some helpful lightning safety tips from the National Weather Service, along with the percentage of lightning fatalities for various outdoor sports: 40% of lightning deaths during sports occur during soccer games, 27% during golf, 17% while running, 10% during baseball, 3% during football, and the remaining 3% from other sports-related lightning deaths.

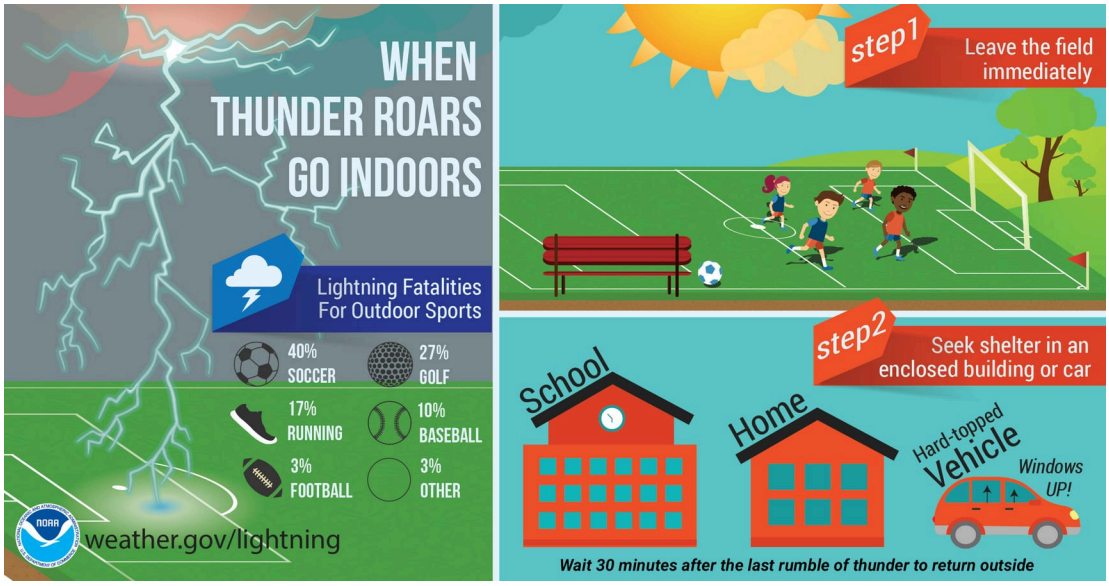



Figure 10.6.1: Lightning safety tips from the National Weather Service. When thunder roars, go indoors and stay there at least 30 minutes after the last clap of thunder. It is also essential to leave the playground immediately and seek shelter in a sturdy structure, such as a school, a house, or even a car. (Public Domain; What constitutes a severe thunderstorm? via NOAA/NWS)

Other safety mottos include "**Turn around, don't drown.**" If you see flowing water over a roadway, don't attempt to drive over it. If there is a possibility of thunderstorms, don't go hiking or do outdoor activities near streams or rivers. Flash flooding is a possibility. Similarly, avoid outdoor locations, as lightning commonly strikes the tallest object around. In an open field, it may be you. 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.

## Tornado Safety

If a tornado is approaching, the safest place is inside an interior room in your house or another sturdy structure. The primary threat from a tornado is flying debris, so stay away from windows or doors. Figure 10.6.2 illustrates these tips via a National

Weather Service bulletin. NWS suggests that when a tornado warning is issued, "get in, get down, and cover up". You need to get into a sturdy structure with as many walls between you and the outside. If such a shelter is not available, get as low as possible in the building you are in. A tornado shelter or cellar is common in the central plains and is the best place to keep yourself safe during a tornado. Once there, take refuge in a closet or bathtub with a mattress or something to cover you. You can cover yourself with pillows, blankets, or wear a helmet to protect yourself from falling objects.

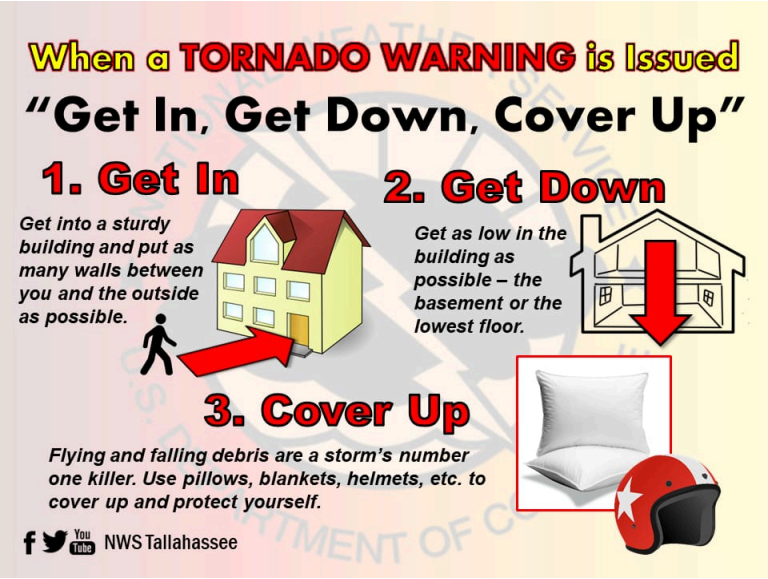



Figure 10.6.2: A tornado safety bulletin issued by the National Weather Service. (Public Domain; Severe Weather Preparedness via NOAA/NWS)

Depending on where you live in the world, some hazards are more likely than others. Still, it is essential to plan in advance what you would do in a severe storm. Often, there isn't much time to react, so planning and preparedness can save lives.

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# 10.7: Exercises

## Chapter 10: Exercises

1. Describe the importance of updraft separation from the downdraft and precipitation portion of a thunderstorm.
2. Fill in the blanks.

**Query 10.7.2**

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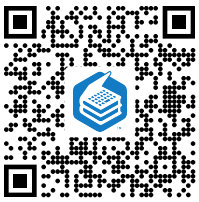


3. Describe the importance of updraft separation from the downdraft and precipitation portion of a thunderstorm.
4. What are the conditions necessary for thunderstorm formation?

**Query 10.7.4**

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5. Describe the importance of wind shear for thunderstorms.
6. Why does hail have concentric layers that alternate clear and opaque ice?
7. Sequence of events.

**Query 10.7.7**

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8. What are the safety precautions you need to take during a thunderstorm?

### Query 10.7.8

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9. *FLASH* You saw a lightning strike! 15 seconds go by... *BOOM* You heard thunder! How far away was the lightning strike?
10. 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?

### Query 10.7.10

#### Interactive Content

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# 10.8: Accessible Descriptions

## Figure 10.1.1: Three Stages of Thunderstorm Development

Three-panel diagram showing the life cycle of a cumulonimbus cloud: towering cumulus stage, mature stage, and dissipating stage. Each panel includes a vertical height scale from the surface up to about 40,000 feet (approximately 12.2 km), with intermediate levels marked at about 10,000 ft (3.0 km), 20,000 ft (6.1 km), and 30,000 ft (9.1 km). A dashed line indicates the freezing level near 32°F (0°C), located around 10,000–15,000 ft (about 3–4.5 km).

In the towering cumulus stage, strong upward motion dominates, with warm air rising from the surface to heights above 20,000 ft (6 km). In the mature stage, the cloud extends to near 40,000 ft (12 km), with both strong updrafts and downdrafts present; precipitation forms and falls, and air circulates within the cloud. In the dissipating stage, downward motion dominates, with downdrafts extending from the upper cloud (around 30,000–40,000 ft or 9–12 km) toward the surface, and the cloud gradually weakens.

The diagram emphasizes that vertical air motion changes from primarily upward (early stage), to mixed upward and downward (mature stage), to primarily downward (dissipating stage), while the cloud reaches its greatest height and intensity during the mature phase.

*Table 10.8.1: Storm stages and characteristics*

| Stage | Height range (ft / km) | Dominant motion | Key processes | Cloud characteristics |
|---|---|---|---|---|
| Towering cumulus | Surface to ~20,000+ ft (0–6+ km) | Updrafts | Rising warm air, condensation | Growing vertical cloud |
| Mature stage | Surface to ~40,000 ft (0–12 km) | Updrafts and downdrafts | Precipitation, strong circulation | Fully developed storm |
| Dissipating stage | Surface to ~30,000+ ft (0–9+ km) | Downdrafts | Rainfall dominates, weakening updrafts | Cloud breaks apart |

*Table 10.8.2: Key atmospheric levels:*

| Feature | Approx. height (ft) | Approx. height (km) | Temperature reference |
|---|---|---|---|
| Freezing level | ~10,000–15,000 | ~3–4.5 | 0°C (32°F) |
| Mid-cloud region | ~20,000 | ~6.1 | Below freezing |
| Upper cloud top | ~30,000–40,000 | ~9–12 | Much colder |

## Figure 10.1.2: Single Celled Thunderstorm Life stages

Wide diagram showing the typical lifecycle of a single-cell thunderstorm over a time span of about 30 minutes, progressing from left to right through cumulus, developing, mature, and dissipating stages. A dashed horizontal line marks the freezing level at 0°C, separating liquid droplets below from ice and snow above.

In the cumulus stage (around 0 minutes), small clouds form with rising air and droplets below the freezing level. In the developing stage, stronger updrafts carry moisture upward above the freezing level, forming snow and ice aloft; moderate precipitation may begin. In the mature stage (roughly mid-lifecycle), the cloud reaches its greatest vertical extent, with strong updrafts and downdrafts occurring simultaneously. Precipitation is heaviest at this stage, including rain below the 0°C level and snow or ice above, and lightning may occur. In the dissipating stage (approaching ~30 minutes), downdrafts dominate, precipitation weakens to light rain or snow, and the cloud gradually spreads and thins.

The diagram shows vertical motions with arrows: upward motion (updrafts) dominating early, both upward and downward motion

in the mature phase, and mainly downward motion (downdrafts) in the final stage. Ice is present at higher altitudes above the freezing level, while liquid droplets dominate below.

## Figure 10.1.3: Squall line of Thunderstorms

Weather radar image showing precipitation intensity (reflectivity, measured in dBZ) across parts of the central and eastern United States, including Illinois, Indiana, Kentucky, and Ohio. The color scale ranges from about 10 dBZ (very light precipitation) to over 70–80 dBZ (extreme intensity). A narrow, continuous line of strong storms extends from southwest to northeast, roughly from western Kentucky near Murray and Princeton through Evansville and Terre Haute into Indiana. Along this line, reflectivity values reach approximately 50–70+ dBZ, indicating heavy rain and potentially severe thunderstorms.

Behind the line (to the west), a broad area of lighter to moderate precipitation is present, with reflectivity values around 20–40 dBZ. Ahead of the line (to the east), much of Kentucky and southern Indiana shows little or no precipitation, except for a large area—of weak radar returns around Louisville and surrounding regions, generally below about 10–30 dBZ, indicating light precipitation or possibly non-precipitation echoes. Isolated storm cells are visible farther northeast in Indiana and Ohio, with localized reflectivity values around 40–60 dBZ.

Overall, the image shows a squall line with the most intense precipitation along the leading edge, lighter precipitation trailing behind, and mostly clearer conditions ahead of the line.

## Figure 10.1.4: Mesoscale Convective System

Weather radar image showing a large, organized mesoscale convective system (MCS) with a curved, rotating appearance. Reflectivity (in dBZ) ranges from about 10 dBZ (very light precipitation) to over 60–70 dBZ (very intense rainfall). Near the center of the system is a region of lower reflectivity (around 0–20 dBZ), not a true eye but an area of weaker precipitation within the storm. Surrounding this region is a broad zone of intense convection, with reflectivity values around 50–70+ dBZ, indicating heavy rain and strong thunderstorms.

Bands and clusters of storms extend outward from the center—especially on one side, with reflectivity values generally between 30 and 60 dBZ, representing moderate to heavy rainfall. Outer areas of the system show lighter precipitation, with values around 10–30 dBZ. Coastlines are visible near the storm, indicating the system is close to land, but the strongest precipitation is concentrated within the convective core and organized bands.

Overall, the image shows a mesoscale convective system with a central region of weaker echoes, surrounded by intense convective cells and organized rainbands, rather than the symmetric structure and clear eye of a tropical cyclone.

## Figure 10.1.5: Supercell Thunderstorm Diagram

Diagram of a supercell thunderstorm showing its main structural features from southwest (SW) to northeast (NE). A tall cumulonimbus cloud extends vertically to the upper atmosphere, spreading outward into an anvil near the top, typically around 10 to 12 kilometers in height. Above the anvil, a localized dome called an overshooting top rises slightly higher, indicating strong upward motion. Beneath the anvil on one side are mammatus clouds, which appear as downward bulges associated with turbulent air.

At mid-levels, the main storm tower (cumulonimbus) contains strong updrafts, while a line of smaller growing clouds (flanking line) is located on the southwestern side. Near the surface, the cloud base is uneven, with visible striations indicating rotation. A localized lowering called a wall cloud forms beneath the main updraft region, typically a few hundred meters to about 1 kilometer above the ground. Adjacent to this is a precipitation-free base, while farther northeast a shelf cloud marks the leading edge of cooler outflow air. Beneath the shelf cloud is a region of precipitation, where rain and possibly hail fall to the ground.

Overall, the diagram shows a vertically developed storm with key features including an anvil at about 10 to 12 km, a wall cloud near the surface below 1 km, and precipitation concentrated on the forward (northeast) side.

# Figure 10.1.6: Real-life Supercell Thunderstorm

Photograph of a supercell thunderstorm showing a large cumulonimbus cloud with distinct structural features. The cloud extends vertically from near the ground to the upper atmosphere, reaching approximately 10 to 12 kilometers in height. At the top, a flattened anvil spreads outward, with an overshooting top rising slightly higher (around 12 to 13 km), indicating a strong updraft.

On the left side, a flanking line of smaller developing clouds is visible, feeding into the main storm. Near the base of the storm, a lowered region called a wall cloud forms at roughly 0.5 to 1 kilometer above the ground. Beneath this wall cloud, a tornado is present, extending from the cloud base to the surface. To the right of the storm, a region of precipitation is visible, where rain and possibly hail are falling from mid-levels (approximately 2 to 8 km) to the ground.

The image shows a classic supercell structure, with strong vertical development, organized inflow from the flanking line, a rotating wall cloud, and severe weather including a tornado and heavy precipitation.

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# 11: ATMOSPHERIC OPTICS

## Introduction



Figure 11.1: Rainbow over the Ala Wai Canal in Hawaii. (CC-BY-SA 4.0, Shintaro Russell via Atmospheric Processes and Phenomena).

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 the interaction 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.

### Learning Objectives

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

- Explain why the sky is blue, and clouds are white
- Describe how one can observe a rainbow
- Determine from visibility whether objects are close or far away


11.1: Visible Light and Scattering

11.2: Visibility

11.3: Exercises for Chapter 11

11.4: Accessible Descriptions



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# 11.1: Visible Light and Scattering

## The Visible Spectrum

Recall the electromagnetic energy spectrum from Chapter 2, with the visible portion near the center between ultraviolet and infrared:

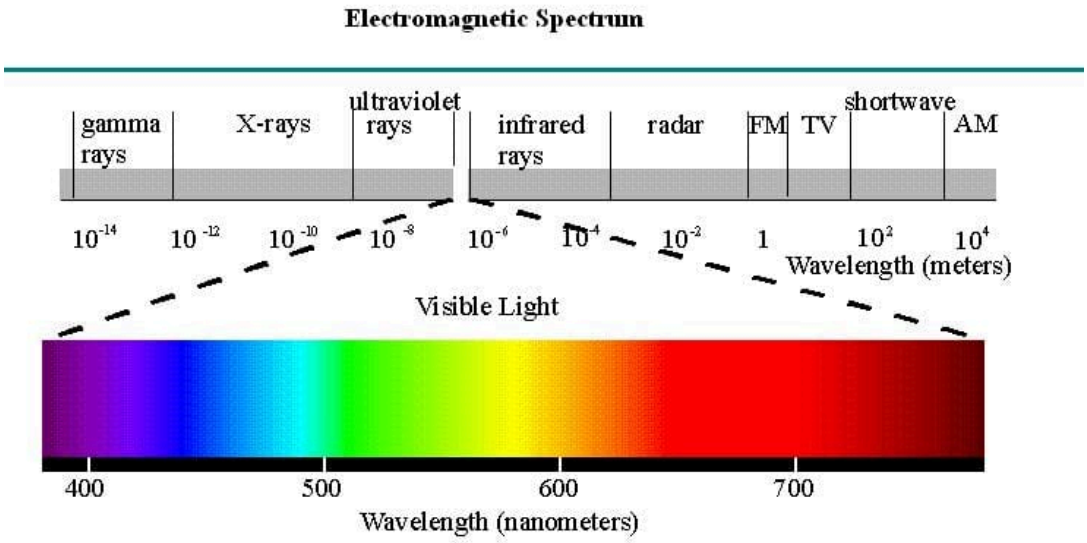



Figure 11.1.1: The electromagnetic energy spectrum highlights the wavelength of visible light. The visible spectrum spans wavelengths from approximately 400 to 700 nm. The shortest wavelengths appear violet and blue, while the longest wavelengths appear red. Wavelengths from 500 to 600 nm appear green and yellow/orange, respectively. (CC-BY-NC-ND 2.0; Scfiasco via Flickr).

Figure 11.1.1 is a diagram of the electromagnetic spectrum showing major categories of electromagnetic radiation arranged by wavelength. The diagram includes gamma rays, X-rays, ultraviolet radiation, visible light, infrared radiation, radar, and radio waves. Wavelength values are shown on a scale in meters, ranging from extremely small to very large. A magnified section highlights the visible-light region of the electromagnetic spectrum, indicating that it occupies a narrow range within the broader spectrum and is typically measured in nanometers, from 400 to 700 nm. A breakdown of colors by wavelength is provided in Table 11.1.1.

*Table 11.1.1: Visible Light Wavelength Ranges*

| Perceived Color | Approximate Wavelength Range (nanometers) |
|---|---|
| Violet | 380–450 nm |
| Blue | 450–495 nm |
| Green | 495–570 nm |
| Yellow | 570–590 nm |
| Orange | 590–620 nm |
| Red | 620–750 nm |

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 significantly alters how light interacts with particles in the atmosphere. Before we begin, let's define a few key terms to help us understand the material that follows.

1. White light contains **all** colors. Sunlight is considered white light.
2. The color an object appears to be is due to the reflection of that color toward your eye. For example, a green leaf looks green

because it absorbs red and blue light but reflects green light.

Based on this simple description, we can say that the sky appears blue because molecules in the atmosphere reflect blue light. But why do they reflect blue light more than red light? This concerns the relative sizes of the molecules and the wavelengths of light that interact with them. In general, there are three laws governing the interactions between wavelength and 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 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 of 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 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 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 the fourth power of the wavelength of light.

$$S \sim \frac{1}{\lambda^4}$$

Because this relationship is inversely proportional to wavelength (λ), shorter wavelengths are scattered more strongly. 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.

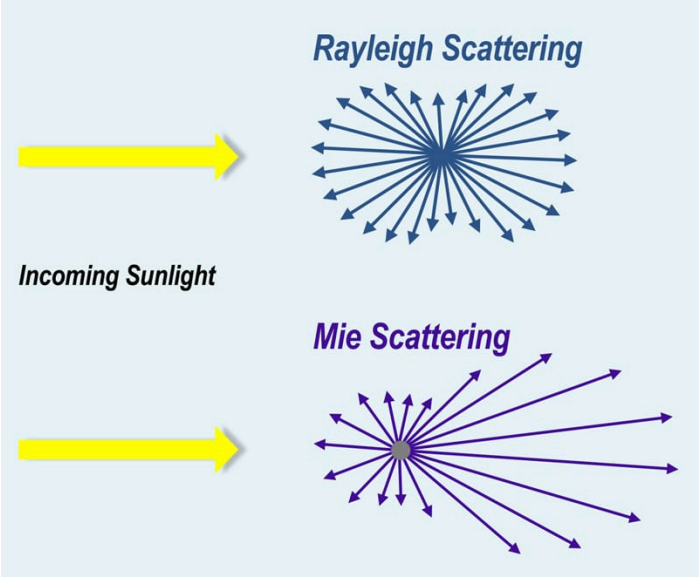



Figure 11.1.2: During Rayleigh scattering, light is scattered close to uniformly off the object in all directions. Meanwhile, Mie scattering scatters light more strongly in the forward direction and exhibits a weaker wavelength dependence. (Public Domain; Shedding Light on PACE via NASA)

## Mie Scattering

Mie scattering, on the other hand, is the reason clouds appear 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 illuminating color is the same as the reflecting color. In the case of clouds, they appear white because white light from the sun illuminates them. Figure 11.1.2 shows the difference between Rayleigh and Mie scattering.

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# 11.2: Visibility

## Geometric Optics

Finally, geometric optics explains rainbows by tracing a beam of light, called a ray. Unlike scattering, geometric optics is used only when the size of the objects interacting with light is orders of magnitude larger than the wavelength, e.g., raindrops. Tracing a ray of light as it passes through a raindrop helps us understand how and why rainbows form. When light encounters an interface between two media, such as air and water, it can reflect or refract. Refraction is responsible for everyday conundrums, as in Figure 11.2.1.



Figure 11.2.1: Glass filled with water placed on a flat surface outdoors. A pen is partially immersed in the water at an angle. The portion of the rod submerged in water appears displaced relative to the portion above the surface, illustrating how light refracts when it passes from air into water. (CC-BY-NC 2.0, Martin LaBar via Flickr).

Refraction occurs when rays bend at an interface between two media. Reflection occurs when rays bounce back from an object. These two phenomena are responsible for atmospheric optical phenomena such as rainbows, halos, and sundogs. Here, we focus solely on rainbows, the geometric optics observed in liquid raindrops.

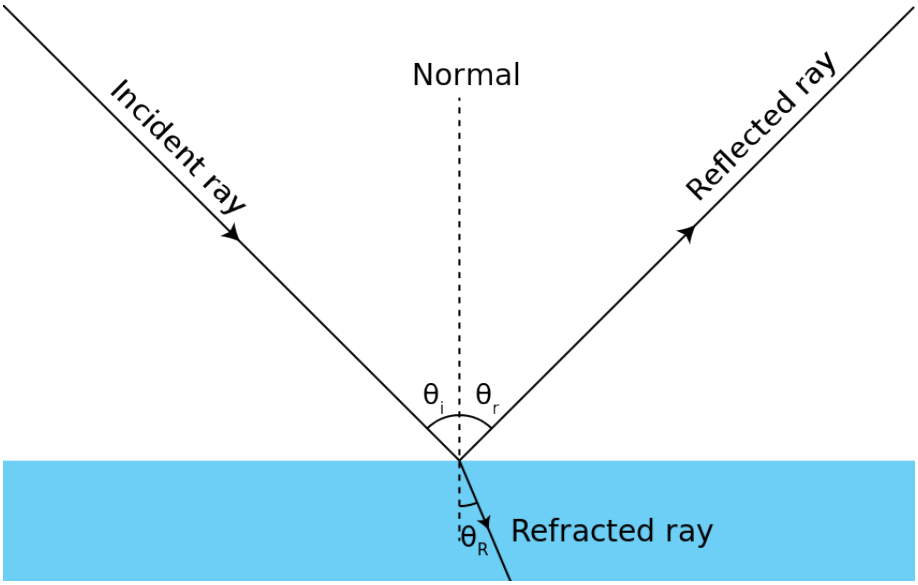



Figure 11.2.2: Diagram showing angle of incidence, reflection, and refraction of a ray of light. (Public Domain; Nilock, via Wikimedia Commons). Alternative Description of Image.

Figure 11.2.2 shows a light ray incident on an interface between two media with different refractive indices, such as air and water. In this case, some of the incoming (incident) light will either refract from the air into the water, reflect back into the air, or be absorbed into heat. In the case of light interacting with a raindrop, we observe 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 Figure 11.1.3 shows the interaction of light with a raindrop 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 split the colors of light, as a prism does. This is because the refractive index of light depends on wavelength; shorter wavelengths refract more strongly than longer wavelengths.

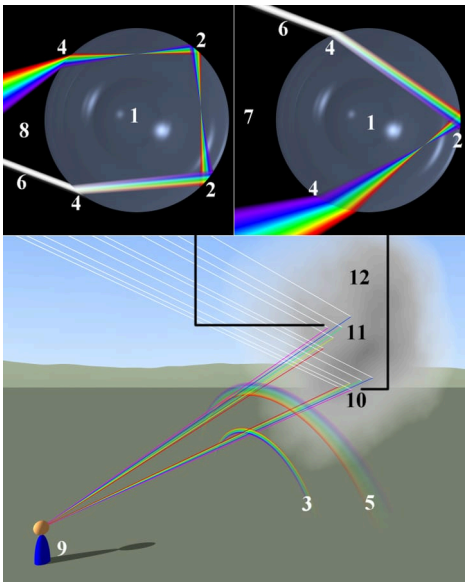



Figure 11.2.3: The formation of a rainbow with light propagation. Legend: 1.) Spherical droplet 2.) Places where internal reflection of the light occurs 3.) Primary rainbow 4.) Places where refraction of the light occurs 5.) Secondary rainbow 6.) Incoming beams of white light 7.) Path of light contributing to the 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; Peo, via Wikimedia Commons). Alternative Description of Image.

Put concisely, a rainbow is an atmospheric phenomenon displaying a spectrum of light as a result of the refraction of light in

water droplets. This is possible because the wavelength of light is much smaller than the size of the raindrop. 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 is shown in Figure 11.1.3. At a given time, rainbows appear different to different observers because of different interactions between incoming light rays and water droplets. For instance, one observer may see a well-defined rainbow in an area of large raindrops and bright sun. Another observer half a kilometer away with smaller raindrops 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; secondary rainbows are higher than primary rainbows (Figure 11.1.4). 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.

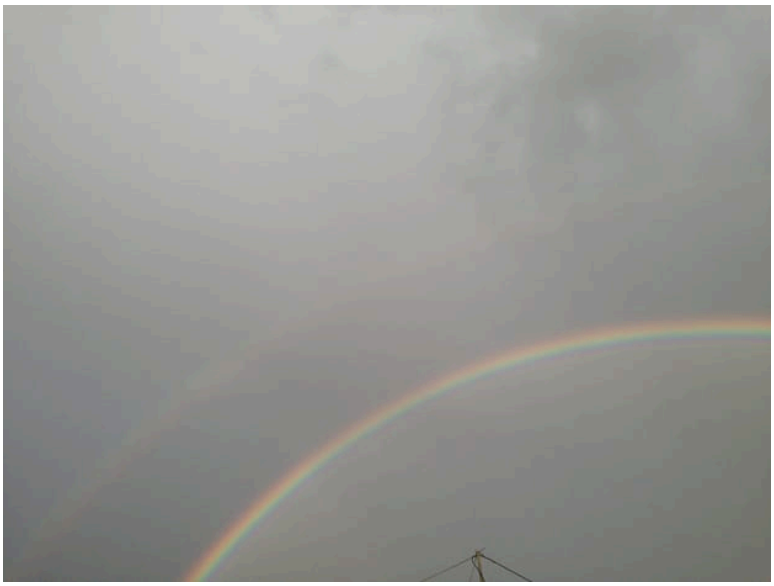


Figure 11.2.4: Double rainbow in a stormy sky. Notice how the colors are flipped in the two rainbows, and how the secondary rainbow is not as bright as the primary rainbow (CC-BY-SA 4.0; Shintaro Russell via Atmospheric Processes and Phenomena).

## 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 behind them and they are facing a region of rain. 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 common in Hawai'i due to abundant sunlight and frequent rainfall.

# 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 hand, 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. Figure 11.2.5 shows the Great Smoky Mountains along the Tennessee–North Carolina border in the southeastern United States. The trees in the foreground are nearly clear, but with each additional ridge, the mountainside appears increasingly white. We assume that each ridge has the same type of foliage as the first. This indicates that objects farther away appear brighter or hazier.



Figure 11.2.5: A photo from the Great Smoky Mountains National Park illustrating that objects far away appear hazier. (CC-BY 2.0; Dave Ynin via Flickr).

The change in brightness is attributable to particulate matter in the atmosphere. This can come in the form of aerosols or hydrometeors (cloud droplets, raindrops, snowflakes, 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 appears white because *Mie scattering* is occurring. Figure 11.2.6 presents an image of the same scene on a clear and polluted day in Sequoia and Kings Canyon National Park.

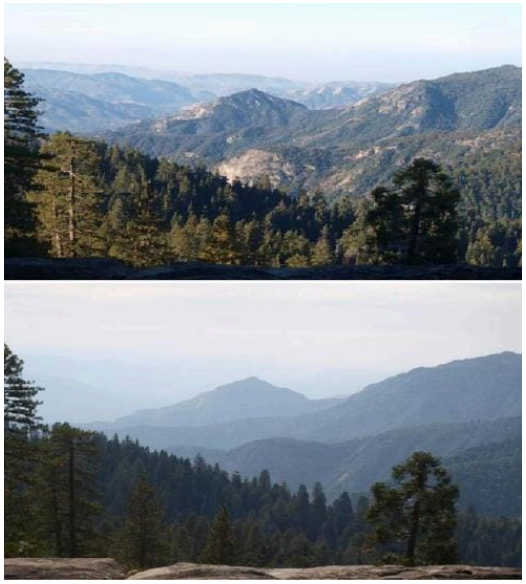


Figure 11.2.6: A clear (top) and polluted (bottom) day at Sequoia & Kings Canyon National Park. (Public Domain; National Park Service).

This national park is often affected by pollution from the Central Valley of 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 in the bottom image than in the top.

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# 11.3: Exercises for Chapter 11

## Exercises

1. Explain the relationship between the diameter of a molecule/object and wavelength of light for:
   a. Rayleigh Scattering
   b. Mie Scattering
   c. Geometric Optics
2. Provide an example of (hint: some are given in the chapter):
   a. Rayleigh Scattering
   b. Mie Scattering
   c. Geometric Optics
3. Why is it not possible to see a rainbow on the ground at noon?
4. What causes low visibility? What type of scattering is this an example of?

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# 11.4: Accessible Descriptions

## Figure 11.2.1: Reflection and Refraction of light

Partial diagram illustrating refraction at a flat boundary between two media. A straight interface separates regions, with a normal line drawn perpendicular at the point where a light ray meets the boundary. The incident ray approaches the boundary and changes direction as it enters the second medium. The transmitted path is labeled “refracted ray,” and the angle between this ray and the normal is marked as $\theta_r$. The layout emphasizes how the ray bends relative to the normal when crossing the boundary.

## Figure 11.2.2: Rainbow Components

Multi-panel diagram explaining how rainbows form through refraction and internal reflection in raindrops.

Top left panel: A spherical raindrop (1) shows a beam of incoming light separated into components (8) entering the drop. At the surface, the light bends (4), travels داخل the drop, reflects off the inner surface, and bends again as it exits at two labeled points (2). The paths illustrate dispersion into different components as the light changes direction at each interaction with the boundary.

Top right panel: A second raindrop (1) shows an alternate path. A beam (6) enters the drop, bends at the surface, reflects once inside (4), and exits at a different location (2). The surrounding region (7) represents the medium outside the drop. This panel emphasizes how varying entry and exit points affect the outgoing light paths.

Bottom panel: An observer (9) stands on the ground while multiple incoming parallel light rays (12) approach and interact with raindrops. The rays bend upon entering and exiting the drops and reflect internally, forming concentrated outgoing paths (10, 11) directed toward the observer. These paths create two arcs: the primary rainbow (3), formed by one internal reflection, and the secondary rainbow (5), formed by two internal reflections. The diagram highlights how specific angles of emerging light relative to the observer produce visible rainbow arcs.

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# 12: GLOBAL CLIMATE AND CLIMATE CHANGE

## Introduction

In the past few chapters, we have examined multiple weather phenomena, including their causes, characteristics, and effects. The weather is the instantaneous state of the atmosphere around us, as depicted in Figure 12.1. It consists of short-term variations, occurring over minutes to days, in variables such as temperature, precipitation, humidity, air pressure, cloudiness, radiation, wind, and visibility. Due to the chaotic, short-term nature of weather, its predictability is limited to days.

Although the weather can be unpredictable, the climate is more predictable. The weather on a particular winter day in San Jose may be colder than on the same day in Lake Tahoe, but, on average, Tahoe's winter climate is significantly colder than San Jose's. Climate, then, is the long-term average of weather. A good climate is why we choose to vacation in Hawaii in February, even though the weather is not guaranteed to be ideal every day.



Figure 12.1: Weather changes from day to day and place to place. (Public Domain; Difference between weather and climate via NCEI/NOAA)

Climate can be thought of as the average weather that varies slowly over decades or centuries. It includes statistics, such as the probabilities or frequencies of extreme weather events. For climate, the states of the atmosphere, ocean, ice, land surface, and biosphere are important. Climate is potentially predictable for extended time periods if the forcing is known, because the principle of energy conservation controls Earth's average temperature. In short: *'Climate is what you expect. Weather is what you get.'*

Thus, weather and climate are related, but they differ in the time scales of changes and their predictability. In this chapter, we'll explore the factors that shape a region's climate and the various climates found worldwide. We'll also analyze factors that may be affecting our climate and how climate change impacts our future.

### Learning Objectives

- Explain the controlling factors of global climate.
- Describe the various global climates and biomes classified using the Köppen Classification System.
- Analyze Earth's climatic past and how scientists measure and collect climate data.
- Describe the causes of short-term and long-term climate change.

12.1: Regional Climates

12.2: Proxies
12.3: Past Climate Changes
12.4: Causes of Climate Change
12.5: Recent Climate Change
12.6: Solutions
12.7: Exercises
12.8: Accessible Descriptions

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- 1.12: Climate- Systems and Change by Cameron McCormick is licensed CC BY-NC 4.0.

# 12.1: Regional Climates

## Controls on Climate

As mentioned earlier, climate is the average weather at a particular location over a long period, typically 30 years or more. Climate can change, but only over long periods of time. The climate of a region depends on its position relative to many things. These factors are described in the following sections. A useful mnemonic device for remembering all the major influences is LACEMOPS. These are some factors influencing the weather, but over a long period:

- **Latitude:** The primary factor influencing a region's climate is its latitude, as different latitudes receive varying amounts of solar radiation. As discussed in Chapter 2, the equator receives the most solar radiation. Days are equally long year-round, and the sun is just about directly overhead at midday. The polar regions, on the other hand, receive the least solar radiation. The night may last nearly six months during the winter, at the highest latitudes. Even in summer, the sun never rises very high in the sky. Sunlight filters through a thick wedge of the atmosphere, making it much less intense. The high albedo, because of ice and snow, reflects a good portion of the sun's light. Thus, the equatorial regions of Earth will experience a warm climate year-round, whereas the higher latitudes will be much colder, with significant seasonal temperature variations.
- **Air Masses:** Recall from Chapter 8 that large bodies of air over some areas of the Earth take on the temperature and moisture characteristics of the surface over which they form. They often end up prescribing the region's climate.
- **Continentality:** When a particular location is near an ocean or large lake, the body of water significantly influences the region's climate. In a maritime climate, temperatures vary relatively little from season to season and from day to day. For a location to have a true maritime climate, the winds must most frequently come off the sea. On the other hand, a *continental climate* is more extreme, with greater temperature differences between day and night and between summer and winter.
- **Elevation:** As altitude increases, temperature generally decreases because the air is thinner and the land is farther from the Earth's surface, which is a source of heat. Locations at higher elevations usually experience lower annual temperatures and drier conditions compared to their sea-level counterparts.
- **Mountain barriers (orographic effect):** Mountain ranges have two effects on the surrounding region's climate. The first is the rain shadow effect, which brings a warm, dry climate to the leeward side of a mountain range. The second effect mountains have on climate systems is the ability to separate coastal regions from the rest of the continent. Since a maritime air mass may have trouble rising over a mountain range, the coastal area will have a maritime climate, but the inland area on the leeward side will have a continental climate.
- **Ocean Currents:** The temperature of the offshore waters influences the temperature at a coastal location, particularly when winds blow off the sea. The cooler waters along the western United States are caused by a clockwise-rotating ocean current that brings cold water from the Arctic toward the equator. The climatic effect is that coastal regions of California, Oregon, and Washington are cool. Coastal upwelling also brings cold, deep water to the ocean surface off California, contributing to the cool coastal temperatures. In the Atlantic Ocean, the northern ocean current, called the *Gulf Stream*, brings warm water from the tropics to the southern states. This is a significant reason the southern states experience humid conditions in the summer and tornadoes: all this warm, moist air. The Gulf Stream also impacts Europe by bringing warm water northward, making this region warmer than expected.
- **Pressure Systems:** High- and low-pressure systems influence wind direction and atmospheric circulation. Recall from Chapter 7 that semi-permanent highs and lows persist in roughly the same locations worldwide. They affect the climate in those locations.
- **Storm Tracks:** The paths that weather systems, such as mid-latitude cyclones and tropical storms, tend to follow are called storm tracks. The location of these tracks is often influenced by prevailing winds, the jet stream, and high- and low-pressure systems.

## Climate Zones and Biomes

A **climate zone** results from the climate conditions of an area: its temperature, humidity, amount and type of precipitation, and the season. A climate zone is reflected in a region's natural vegetation. Perceptive travelers can determine which climate zone they are in by looking at the vegetation, even if the weather is unusual for that climate on that day.

The major factors that influence climate determine the different climate zones. In general, the same climate zone is found at similar latitudes and in similar positions across nearly all continents, in both the Northern and Southern Hemispheres. The one exception to this pattern is the continental climate, which is not found at higher latitudes in the Southern Hemisphere. This is

because the landmasses of the Southern Hemisphere are not wide enough to support a continental climate.

The most common system for classifying climatic zones is the **Köppen classification**, as illustrated in Figure 12.1.1. This system is based on temperature, precipitation amounts, and the times of year when precipitation occurs. Since climate determines the types of vegetation that grow in an area, vegetation is used as an indicator of climate.

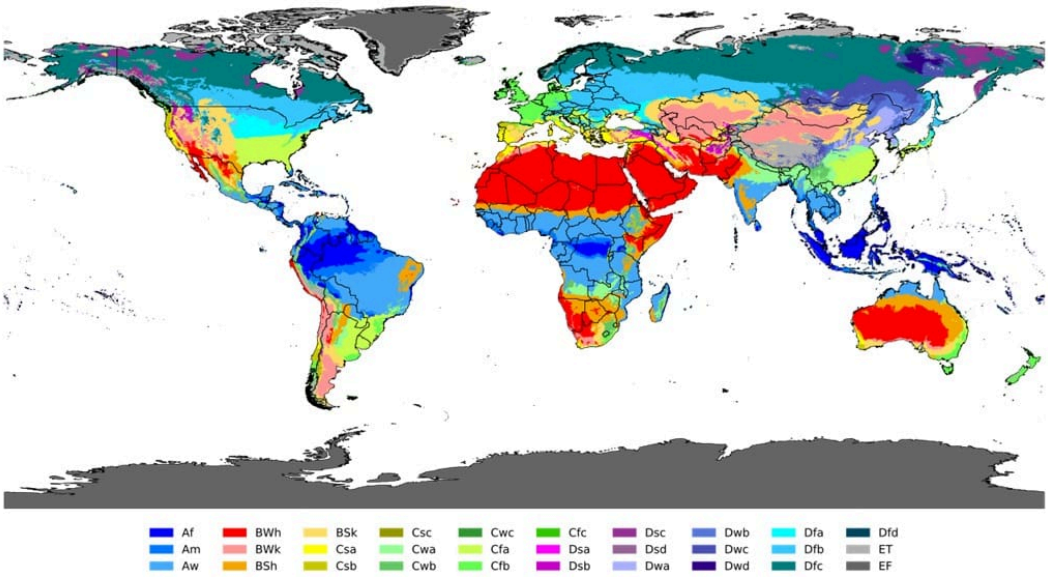



Figure 12.1.1: The Köppen classification system recognizes five major climate groups, each with a distinct capital letter A through E. Each lettered group is divided into subcategories, such as forest (f), monsoon (m), and wet/dry (w), based on precipitation amount and the season in which it occurs. (CC BY 4.0; Beck et al., 2023 via Wikimedia Commons). Alternative description of the image.

A climate type, together with its plants and animals, makes up a **biome**. Organisms across biomes share specific characteristics worldwide because their environments offer similar advantages and challenges. The organisms have adapted to that environment in similar ways over time. For example, different cactus species live on different continents, but they have adapted to harsh desert conditions in similar ways.

## Tropical Moist Climates (Group A)

These climates are found in a band about 15° to 25° N and S of the equator. What climate characteristics are the tropical moist climate group likely to have?

- **Temperature:** Intense sunshine; the average monthly temperature is at least 18°C (64°F).
- **Rainfall:** Abundant, at least 150 cm (59 inches) per year.



Figure 12.1.2: Mt. Kinabalu, Borneo, with the typical vegetation of tropical forests. (CC BY-SA 2.5; NepGrower via Wikimedia Commons)

The wet tropics have almost no annual temperature variation and year-round rainfall of 175-250 cm (65-100 inches). These conditions support the tropical rainforest biome. Densely packed, broadleaf evergreen trees dominate **Tropical Rainforests (Af)**, as depicted in Figure 12.1.2. These rainforests have the highest biodiversity of any ecosystem.

The **Tropical Monsoon (Am)** climate has very low precipitation for 1 to 2 months each year. Rainforests grow here because the dry period is short, and the trees survive on soil moisture. This climate is found where the monsoon winds blow, primarily in southern Asia, western Africa, and northeastern South America**.**

The **Tropical Wet and Dry (Aw)** climate lies between about 5° and 20° latitude, around the Intertropical Convergence Zone (ITCZ). In the summer, when the ITCZ drifts northward, the zone is wet. In the winter, when the ITCZ moves toward the equator, the region is dry. This climate exists where strong monsoon winds blow from land to sea, such as in India. Rainforests cannot survive the months of low rainfall, so the typical vegetation is *savanna***.** This biome consists mostly of grasses, with widely scattered deciduous trees and rare areas of denser forests.

## Dry Climates (Group B)

These climates receive less precipitation than they lose through evaporation. Dry climate zones cover about 26% of the world's land area**.** What climate characteristics are the dry climate group likely to have?

- **Temperature:** Abundant sunshine. Summer temperatures are high; winters are cooler and longer than in Tropical Moist climates
- **Rainfall:** Irregular; several years of drought are often followed by a single year of abundant rainfall

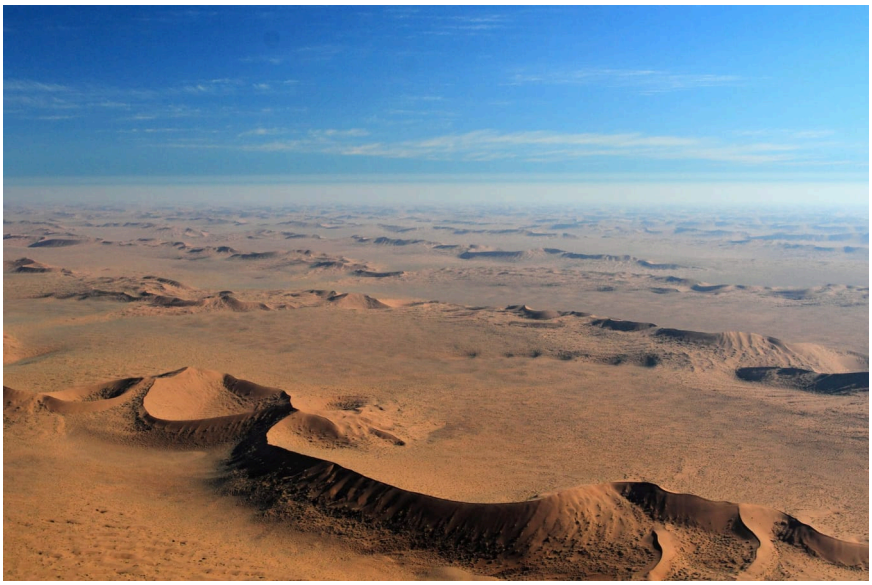


Figure 12.1.3: The Namib Desert in Southern Africa. (CC BY-SA 3.0; Martin Cígler via Wikimedia Commons)

Low-latitude, **Arid Deserts (Bw)** are found between 15° and 30° N and S latitudes, as depicted in Figure 12.1.3. This is where warm, dry air sinks, creating high-pressure zones, as a part of the Hadley Cycle. True deserts make up around 12% of the world's land**.** In the Sonoran Desert of the southwestern United States and northern Mexico, skies are clear. The typical weather is extremely hot summer days and cold winter nights. Although annual rainfall is less than 25 cm (10 inches), rain falls during two seasons. Pacific storms bring winter rains, and monsoons bring summer rains. Since organisms do not have to go too many months without some rain, a unique group of plants and animals can survive in the Sonoran Desert**.**

Higher-latitude **Semi-arid Deserts (Bs)**, also called *steppe*, are found in continental interiors or in rain shadows. Semi-arid deserts receive between 20 and 40 cm (8 to 16 inches) of rain annually. The annual temperature range is large. In the United States, the Great Plains, portions of the southern California coast, and the Great Basin are semi-arid deserts**.**

## Moist Subtropical Mid-latitude (Group C)

A moist subtropical climate is a type of climate found within the temperate or subtropical regions of the Earth. What climate characteristics are the moist subtropical group likely to have?

- **Temperature:** The coldest month ranges from just below freezing to almost balmy, between -3°C and 18°C (27° to 64°F). Summers are mild with average temperatures above 10°C (50°F). Seasons are distinct.
- **Rainfall:** There is plentiful annual rainfall.



Figure 12.1.4: The sage and chaparral vegetation of coastal California. (CC BY-SA 2.0; Jeff Turner via Wikimedia Commons)

The **Dry Summer Subtropical (Cs)** climate is found on the western sides of continents between 30° and 45° latitude. Annual rainfall is 30 to 90 cm (14 to 35 inches), most of which comes in the winter**.** The climate is typical of coastal California, which sits beneath a summertime high-pressure system for about five months each year. Land and sea breezes make winters moderate and summers cool. Vegetation must survive long summer droughts. The scrubby, woody vegetation that thrives in this climate is called *chaparral*, as depicted in Figure 12.1.4.

The **Humid Subtropical (Cfa)** climate zone is found mostly on the eastern sides of continents**.** Rain falls throughout the year, with annual averages of 80-165 cm (31-65 inches). Summer days are humid and hot, ranging from the lower 30s to 40°C (mid-80s to 104°F). Afternoon and evening thunderstorms are common. These conditions are caused by warm tropical air passing over the hot continent. Winters are mild, but middle-latitude storms called cyclones may bring snow and rain. The southeastern United States, with its hot, humid summers and mild, but frosty winters, is typical of this climate zone.

The **Marine West Coast (Cfb)** climate zone extends across western North America between 40° and 65° latitude, an area known as the Pacific Northwest**.** Ocean winds bring mild winters and cool summers. The temperature range, both daily and annually, is fairly small. Rain falls year-round, although summers are drier as the jet stream moves northward. Low clouds, fog, and drizzle are typical. In Western Europe, the climate covers a larger region because there are no high mountains near the coast to block wind blowing off the Atlantic.

## Humid Continental (Group D)

Continental climates are found in most of the North American interior from about 40°N to 70°N. What climate characteristics is the continental group most likely to have?

- **Temperature:** The average temperature of the warmest month is above 10°C (50°F), and the coldest month is below -3°C (-27°F).
- **Precipitation:** Winters are cold and stormy (look at the latitude of this zone and see if you can figure out why). Snowfall is common, and snow can stay on the ground for long periods.



Figure 12.1.5: Fall foliage near Hogback Mountain in Vermont. (CC BY-SA 4.0; chensiyuan via Wikimedia Commons)

Trees grow in continental climates, even though winters are extremely cold, because the average annual temperature is fairly mild. Continental climates are not found in the Southern Hemisphere because of the absence of a continent large enough to generate this effect.

The **Humid Continental (Dfa, Dfb)** climate is found along the polar front in North America and Europe. In the winter, middle-latitude cyclones bring chilly temperatures and snow. In the summer, westerly winds bring continental weather and warm temperatures. The average July temperature is often above 20°C (70°F). The region is typified by deciduous trees, which protect themselves in winter by losing their leaves in the fall, leading to vibrant fall colors as depicted in Figure 12.1.5.

The two variations of this climate are based on summer temperatures:

- Dfa, long, hot summers: summer days may be over 38°C (100°F), nights are warm, and the temperature range is large, perhaps as great as 31°C (56°F). The long summers and high humidity foster plant growth**.**
- Dfb, long, cool summers: summer temperatures and humidity are lower. Winter temperatures are below -18°C (0°F) for long periods.

The **Subpolar (Dfc)** climate is dominated by continental polar air masses that cover the frigid continent. Snowfall is light, but cold temperatures keep snow on the ground for months. Most of the approximately 50 cm (20 inches) of annual precipitation falls during summer cyclonic storms. The angle of the Sun's rays is low, but the Sun is visible in the sky for most or all of the day during the summer, so temperatures may get warm, but are rarely hot. These continental regions have extreme annual temperature ranges. The boreal conifer forests found in subpolar climates are called *taiga* and are characterized by small, hardy, widely spaced trees. Taiga vast forests stretch across Eurasia and North America.

## Polar Climates (Group E)

Polar climates are found across the continents that border the Arctic Ocean, Greenland, and Antarctica. What climate characteristics is the polar climate group most likely to have?

- Temperature: Winters are entirely dark and bitterly cold. Summer days are long, but the sun is low on the horizon, so summers are cool. The average temperature of the warmest month is less than 10°C (50°F). The annual temperature range is extensive.
- Precipitation: The region is dry, with less than 25 cm (10 inches) of rainfall annually; most falls during the summer.



Figure 12.1.6: The shrubs from Denali National Park in Alaska. (Public Domain; Spruce Peak via National Park Service)

The **Polar Tundra (ET)** climate is continental, with severe winters. Temperatures are so cold that a layer of permanently frozen ground, called *permafrost,* forms below the surface. This frozen layer can extend hundreds of meters deep. The average temperature of the warmest months is above freezing, so summer temperatures defrost the uppermost portion of the permafrost. In winter, the permafrost prevents water from draining downward. In summer, the ground is swampy. Although precipitation is low enough in many places to qualify as a desert, evaporation rates are also low, so the landscape receives more usable water than a desert would. Because of the lack of ice-free land near the South Pole, there is very little tundra in the Southern Hemisphere. The only plants that can survive the harsh winters and soggy summers are small, ground-hugging plants like mosses and lichens, small shrubs, and scattered small trees that make up the tundra, as depicted in Figure 12.1.6.

**Ice caps (EC)** are found mainly on Greenland and Antarctica, covering about 9% of the Earth's land area**.** Ice caps may be thousands of meters thick. Ice cap areas have extremely low average annual temperatures, e.g., -29°C (-20°F) at Eismite, Greenland. Precipitation is low because the air is too cold to hold much moisture. Snow occasionally falls in the summer.

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# 12.2: Proxies

## Climate Change in Earth's History

The Earth's climate is the average climate conditions of the Earth as a whole, accounting for all regional variations. For the past two centuries, our global climate has been relatively stable. People placed their farms and cities in favorable climates without considering that the climate could change. But the climate has changed throughout Earth's history, and a stable environment is not the norm.

Measurements with modern instruments (the instrumental record) are available only for roughly the past century. This is insufficient to capture the full natural variability of the climate system, making it difficult to *attribute* current observed changes. We seek to determine whether the recent changes are unusual relative to pre-industrial climate variability. If they are, they are more likely to be anthropogenic. If not, they could well be natural. Paleoclimate research is important for a fundamental understanding of how the climate system works. Some paleoclimate changes, e.g., ice-age cycles, were much larger than those during the instrumental record. Thus, we can learn much from paleoclimate data about the impacts of significant climate changes.

## Methods

Paleoclimate research can extend the instrumental record much further back in time and has delivered a fascinating history of past climate changes. Most paleoclimate evidence is indirect and based on *proxies* for climate variables. This evidence is less precise than measurements from modern instruments because of additional uncertainty in the relationship between the proxy climate variable and the climate system. Examples of proxies include pollen found in lake sediments, which can be used to reconstruct past vegetation cover and, in turn, relate it to temperature and precipitation. Fossil shells of planktic foraminifera, like those depicted in Figure 12.2.1, can be retrieved with a coring device employed from a research vessel. Shells found deeper in the sediment are older. If shells of cold-loving foraminifera are found at a site where currently warm-loving species live, it suggests that near-surface temperatures in the past have been colder. Mathematical methods have been developed to quantify the temperature changes from the species composition. Other proxies are chemical, such as the magnesium-to-calcium ratio (Mg/Ca), which is related to temperature, or oxygen and carbon isotopes in the calcium carbonate shells of foraminifera, which can be used to reconstruct temperature, salinity, ice volume, and carbon cycling. Benthic foraminifera live on or in the ocean's sediments and thus provide useful information on deep ocean properties.

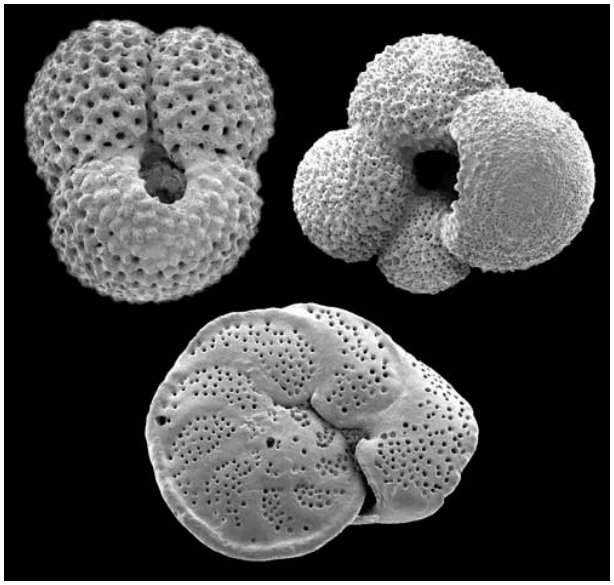


Figure 12.2.1: Electron microscope pictures of Planktic and Benthic foraminifera. (CC BY-NC 4.0; Fossils via Teara)

Proxies are found in different archives such as tree rings, ice cores, corals, ocean or lake sediment cores that cover different time periods at a range of temporal resolutions, as shown in Fig. 12.2.2. Written historical accounts can be used to reconstruct past

climatic conditions at very high temporal resolution – some ancient documents contain daily weather entries – back to about 1,000 years. Still, only a limited number of such records are available. Tree rings, corals, and speleothems (cave deposits such as stalactites and stalagmites) provide reconstructions at annual to decadal resolution back many thousands of years. Ice cores typically have decadal to centennial resolution, extending back almost a million years in Antarctica and about 100,000 years in Greenland. Ocean sediment cores span millions of years but are usually sampled at low temporal resolution, on centennial- to millennial-scale timescales.

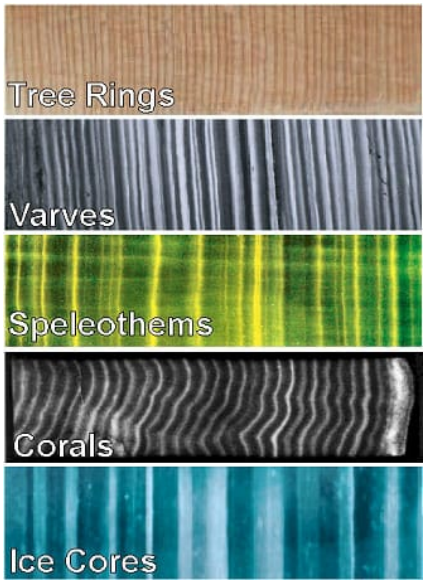



Figure 12.2.2: Layered paleoclimate archives. Tree rings and corals can show annual growth rate depending on local temperature and moisture. Varves are layered sediments from lakes or seas, with light-colored layers from summer and dark-colored layers from winter. Speleotherms include cave formations such as stalactites and stalagmites, which exhibit distinct growth layers. Ice cores also provide a summer-winter color contrast similar to Speleotherms, along with trapped air bubbles detailing past atmospheric composition. (CC BY-NC-SA 4.0; Introduction to Climate Science via Oregon State). Alternative Description of Image.

Several methods are used to date samples to construct chronologies of paleoclimate records. Tree rings are annual layers, which can be counted. Patterns of thin and thick rings can be matched from one tree to another, as illustrated in Fig. 12.2.3. This way, a large number of trees can be used to create a long, layer-counted chronology. Layer counting can also be used in other archives with annual layers, such as ice cores or lake sediments. Most ocean sediments don't have annual layers because of bioturbation, which is the mixing of sediments by worms and other organisms that live in the sediment.

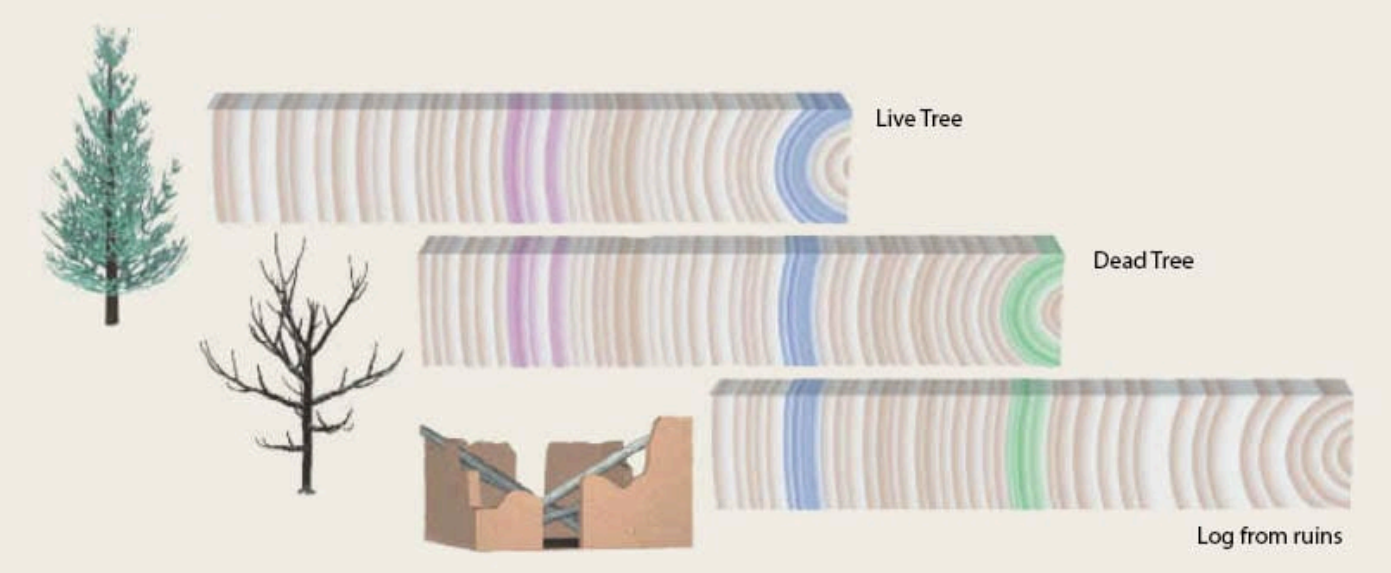



Figure 12.2.3: Climate patterns such as droughts, excessive rainfall, changes in sunlight, etc., result in certain patterns in tree trunks. Such patterns, shown in different colors, will occur at different locations in tree logs from a live tree, a recently dead tree, and an old, logged tree. Thus, long tree-ring chronologies can be constructed by matching overlapping patterns among different trees. (CC BY-NC-SA 4.0; Introduction to Climate Science via Oregon State). Alternative Description of Image.

When organic material is present, radiocarbon dating can be used to determine a sample's age. Radiocarbon ($^{14}C$) decays exponentially with a half-life of 5,730 years. Thus, the lower the radiocarbon-to-carbon-12 ratio ($^{14}C/^{12}C$) in a sample, the older it is. This ratio can be measured precisely with a mass spectrometer. However, this method can be used only until about 40,000 years before the present, because older material contains unmeasurably small amounts of $^{14}C$.

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# 12.3: Past Climate Changes

## The Last Two Millennia

Using proxies and historical accounts, such as pictures of the frozen Thames, as illustrated in Fig. 12.3.1, scientists have obtained a historical record of Earth's climate. Over the last two millennia, proxy temperature and carbon dioxide measurements can be correlated with records of human art and architecture. The PAGES 2k initiative was a global project to reconstruct temperature and hydroclimate records for the last 2,000 years, using data from a wide range of sources like tree rings, ice cores, and corals.



A



B

Figure 12.3.1: A: A famous painting of a Frost Fair on the frozen River Thames around 1683-84 AD. (Public Domain; Thomas Wyke via Wikimedia Commons). B: Ruins of Hvalsey Church from the Greenland settlements of the Norse. (Public Domain; Number 57 via Wikimedia Commons)

The reconstruction suggested that a short period of relatively cold conditions occurred in Europe from the 16th to the 19th centuries, known as the Little Ice Age. Conversely, relatively warm conditions during the $9^{th}$ to $13^{th}$ centuries, known as the Medieval Warm Period, may have enabled Vikings to colonize Greenland and reach North America. Some reconstructions of global temperatures, however, indicate that the Medieval Warm Period was not a global phenomenon, as shown in Fig. 12.3.2. These reconstructions also suggest a long-term cooling trend over the past 2,000 years that culminated in the Little Ice Age, followed by relatively rapid warming in the $20^{th}$ century. According to the PAGES 2k reconstruction, the global average temperature during the 30 years from 1971 to 2000 was warmer than during any other 30-year period over the last 1,400 years. This suggests that the recent warming is unusual. The rate of change over the last ~100 years also seems unusually fast compared with the previous 2,000 years. The two independent reconstructions agree well in the cooling trend over the past 1,000 years, but the PAGES 2k reconstruction suggests slightly warmer conditions during the first millennium CE (Common Era). Another recent attempt to reconstruct past temperatures is the Marcott et al. (2013) dataset, which is based primarily on lower-resolution ocean sediment cores. It is therefore smoother than the higher-resolution PAGES 2k dataset, which includes land-based data such as pollen and tree rings.

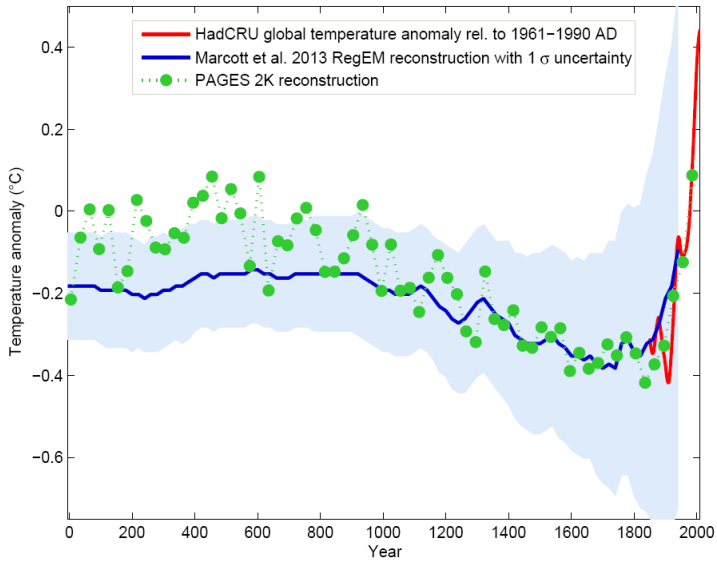



Figure 12.3.2: Global multi-proxy temperature reconstructions from the PAGES 2k project based on seven continental-scale regional reconstructions (green) compared with an independent reconstruction that includes the whole Holocene and the instrumental record (red). The PAGES 2k reconstruction represents 30-year averages. It contains many more data points than the lower-resolution Marcott et al. (2013) reconstruction, particularly for the last 1,000 years. Thus, the error range of the PAGES 2k reconstruction is presumably much smaller than that indicated by the blue shading. (CC-BY-4.0; Klaus Bittermann via RealClimate). Alternative Text Description.

## The Holocene

Fig. 12.3.3 shows the whole Holocene (the last 10,000 years) reconstruction of global average temperatures from the Marcott dataset. It suggests that the long-term cooling trend over the last 2,000 years is part of a longer trend that extends back to the middle Holocene around 4,000 BC. The early Holocene, from around 8,000 BC to 4,000 BC, was relatively warm, similar to recent decades. This remains debated in the scientific community, with recent research suggesting that it wasn't warmer during the early Holocene and that biases in seasonally sensitive proxies are to blame. If this is true, the current warming will be unprecedented for more than 10,000 years, perhaps more than 100,000 years or longer. The rate of temperature change appears to be much slower than it has been over the last 100 years. Still, the reconstruction's relatively low resolution smoothens it and prevents a fair comparison with the instrumental record on 100-year timescales.

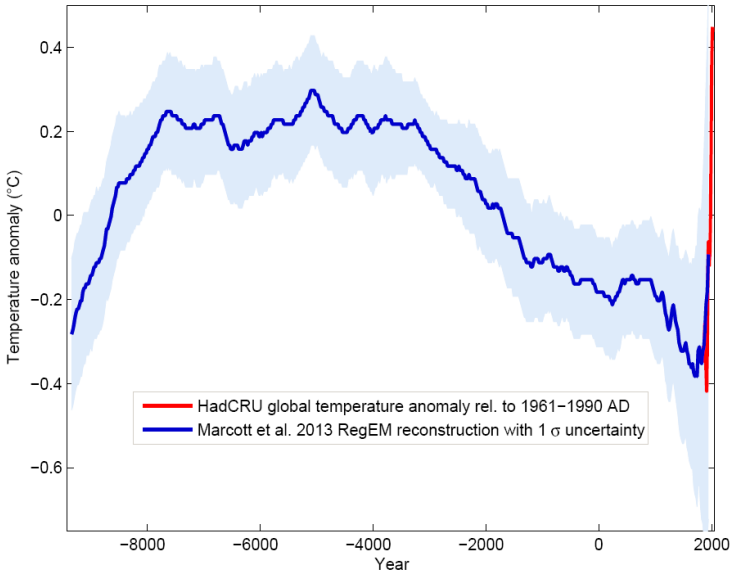



Figure 12.3.3: Holocene global surface temperature reconstruction from Marcott et al. (2013, blue) with shaded error ranges together with the instrumental record (red) as a function of time in years CE. (CC-BY-4.0; Klaus Bittermann via RealClimate). Alternative Text Description.

Now let's examine $CO_2$. Is the observed increase in atmospheric $CO_2$ during the last 60 years unusual compared to the pre-industrial Holocene? Ice cores can be used to answer this question. When snow accumulates on an ice sheet, it is compressed into ice by the overlying snow. During this compaction process, small air bubbles are trapped within the ice, as depicted in Fig. 12.3.4. In the lab, this ancient air can be extracted from the ice by mechanical crushing, and its $CO_2$ concentration, as well as those of other greenhouse gases, can be measured.

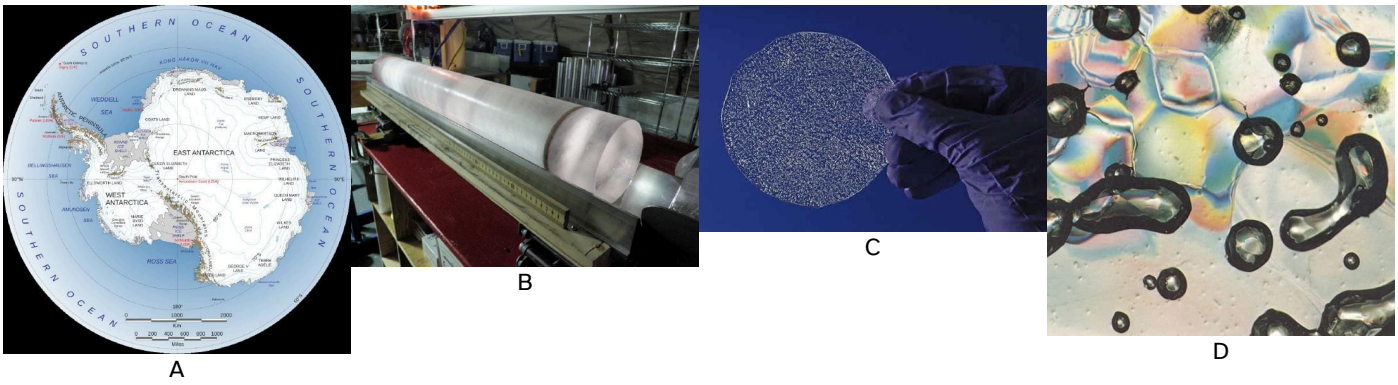



Figure 12.3.4: A: Ice cores have been drilled in different locations in Antarctica by research scientists (Public Domain; NASA via Wikimedia Commons). B: These ice cores are preserved and analyzed to obtain past climate data (Public Domain; Heidi Roop via National Science Foundation). C: When snow falls, tiny bubbles of air get trapped in the ice through compaction of snow. These air bubbles in the ice are visible to the naked eye (CC BY-NC 4.0; R Mulvaney via British Antarctic Survey). E: The air bubbles under a microscope (CC BY-NC 4.0; W Berner via Nature/University of Bern)

Ice cores from Greenland are less suitable for CO2 reconstructions because they are often contaminated with impurities (e.g., dust), which can lead to CO2 production within the ice. However, Antarctic ice is so pure that it provides excellent records of past atmospheric $CO_2$ concentrations. Multiple ice cores have been drilled across Antarctica to study past climate. The measurements from the youngest ice match up very well with the direct measurements of modern air from Mauna Loa, as illustrated in Fig. 12.3.5. Also, the measurements from different ice cores agree with each other (different colored symbols in Fig. 12.3.5). This indicates that Antarctic ice cores faithfully record past atmospheric $CO_2$ concentrations. The results show that atmospheric $CO_2$ concentrations have been relatively constant at about 260-280 ppm during the Holocene. It was only during the last 200 years that $CO_2$ concentrations started to increase. Thus, we conclude that the increase in CO2 over the last 200 years is unprecedented and has not occurred during the preceding 10,000 years. We also know that burning fossil fuels increased dramatically after the Industrial Revolution (1760-1840).

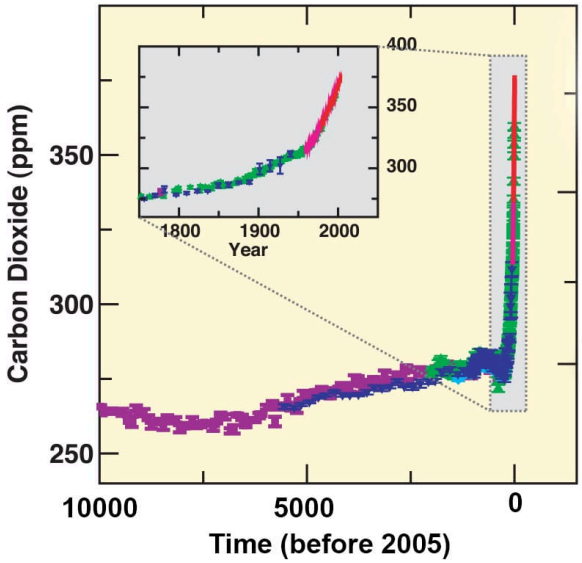



Figure 12.3.5: Ice core $CO_2$ measurements of ancient air. The different colors indicate different ice cores. The inset zooms in on the last 200 years and includes, in red, the modern air measurements from Mauna Loa. Note that the ice core measurements agree well with the modern data where they overlap. (Public Domain; Summary for Policymakers via IPCC 2007). Alternative Text Description.

Other greenhouse gases have also been measured in air extracted from ice cores. Methane ($CH_4$) and nitrous oxide ($N_2O$) show a very similar behavior to $CO_2$, such that their concentrations were relatively constant throughout the Holocene, around 700 ppb and 260 ppb, respectively, and increased dramatically during the last 200 years to values around 1,700 and 310 ppb, respectively (IPCC, 2007).

## The Ice Ages

Fig. 12.3.3 already hints at a cold period before the Holocene. Indeed, we now know that Earth had been in an ice age, or glacial state, for a long time before the current Holocene warm period began. But it was only in the 19$^{th}$ century that scientists realized that Earth had experienced global ice ages. This discovery was made by Louis Agassiz, a Swiss geologist, who hypothesized that not only Alpine glaciers were advanced but also large ice sheets moved south from northern Europe and America, leaving glacial landforms behind, as shown in Fig. 12.3.6.

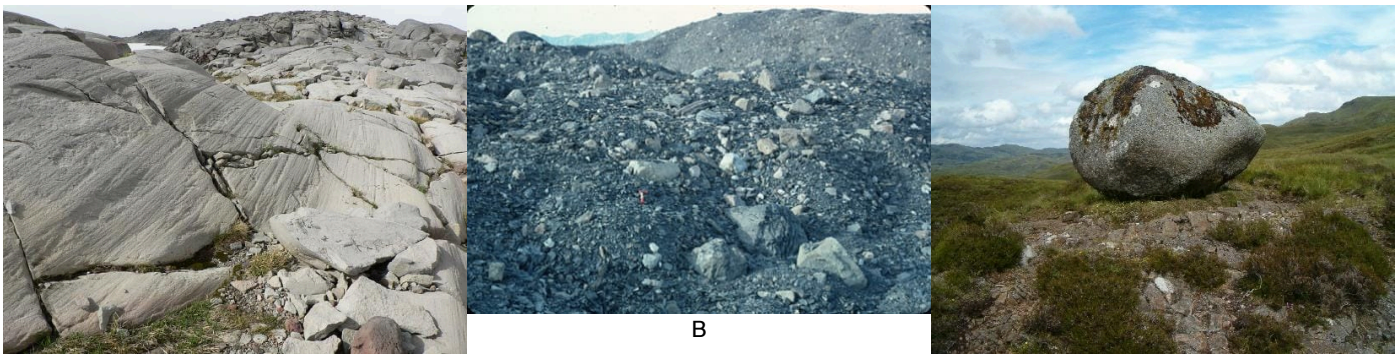



Figure 12.3.6: Glacial landforms across the world provide evidence of past glacial advances. A: Polished bedrock with striations indicates a glacier was moving over it. The glacier incorporates rocks into its base and, by pushing them over the underlying bedrock, creates grooves. This example is from Mount Rainier National Park (CC BY 2.5; Walter Siegmund via Wikimedia Commons). C: Moraines are glacial deposits formed at the side (lateral moraines) or end (terminal moraines) of a moving glacier. This example from Svalbard shows rocks and debris left behind by a glacier (CC BY-NC 4.0; Dr. Karen Kleinspehn via MTU). B: Erratic boulders like this one from Scotland, many miles from a possible bedrock source, have been attributed by Louis Agassiz to the action of ice-age glaciers. (CC BY-SA 2.0; Patrick Mackie via Wikimedia Commons)

During the height of the last ice age, the Last Glacial Maximum (LGM) roughly 20,000 years ago, large additional ice sheets covered parts of North America and northern Europe, as illustrated in Fig. 12.3.7. The Laurentide Ice Sheet was more than 3 km thick and covered all of what is now Canada and part of the northern United States, reaching as far south as New York City, Chicago, and Seattle. The Eurasian (or Fennoscandian) Ice Sheet covered all of Scandinavia, much of the British Isles, the Baltic Sea, and surrounding land areas from northeastern Germany to northwestern Russia. Mountain glaciers also descended farther into valleys and often into lowlands. Because so much more water was locked up as ice on land, the sea level was 120 m lower during the LGM than it is today.

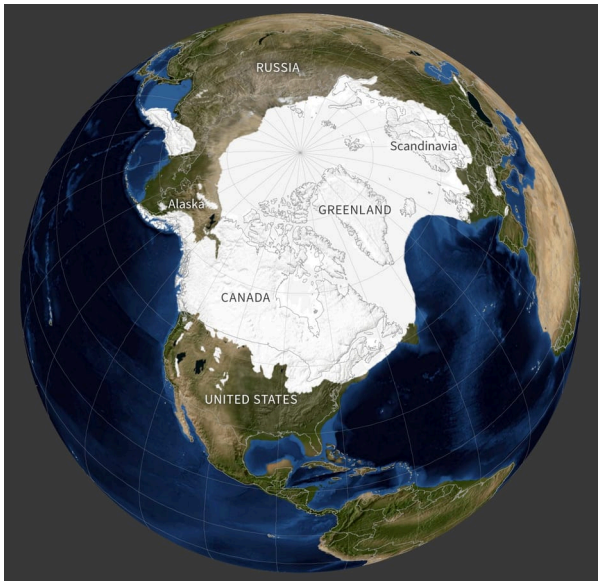



Figure 12.3.7: Reconstruction of ice sheet extent near the peak of the Last Glacial Maximum. (Public Domain; Ice sheet extent via Climate.gov)

The LGM is a well-studied period in paleoclimate research, and a wealth of data is available. Ice cores show lower atmospheric greenhouse gas concentrations, such as near 180 ppm for CO2, compared with 280 ppm during the late pre-anthropogenic Holocene. Vegetation reconstructions indicate that forests were replaced by tundra and grasslands across large areas of the mid- to high latitudes. We also know from ice and ocean sediment cores that the air was dustier. Temperature proxies show colder temperatures almost everywhere. However, temperature changes were not uniform across the region. Temperature changes over large parts of the tropical and subtropical oceans were relatively small. Globally averaged sea surface temperatures have been estimated to be only 2°C cooler than present. Land areas in the tropics experienced moderate cooling of about 3°C. The most significant cooling, exceeding 8°C, occurred over land at mid- to high latitudes and over Antarctica. The globally averaged surface air temperature is estimated to have been 4°C lower during the LGM. Studies also suggest that, on average, land cooling was 3 times greater than ocean cooling. The land-sea contrast and polar amplification are similar to those observed during the past century of warming. This suggests that those are robust properties of the climate system. We conclude that paleoclimate data from the LGM indicate that Earth was dramatically different from today, with large ice sheets, low sea levels, and other vegetation. These changes occurred even though global average temperatures changed by only 4-5°C. This is comparable to changes projected for some future scenarios.

Fig. 12.3.8 shows that there were about nine glacial-interglacial cycles during the past 800,000 years. Most of the time, the sea level was lower than today, reaching more than 120 m below sea level during some glacial maxima. Oxygen isotope ($\delta^{18}O$) measurements in benthic foraminifera from deep-sea sediments indicate the extent to which water was locked up as ice. Higher ice volume will result in lower global sea levels. Ice core records also show that glacial periods were consistently associated with low atmospheric $CO_2$ concentrations and low temperatures in Antarctica. $CO_2$ concentrations varied between about 180 ppm during glacial maxima and 280 ppm during interglacials. The correlation between these completely independent datasets, one from Antarctic ice cores, the other from deep-sea or benthic foraminifera, is astounding. It demonstrates that climate and the carbon cycle are tightly interlinked. High $CO_2$ concentrations are always associated with warm temperatures, high sea levels, and low ice volume. This indicates the importance of atmospheric $CO_2$ concentrations for climate, but it also suggests that climate influences the carbon cycle and, in turn, alters $CO_2$ levels.

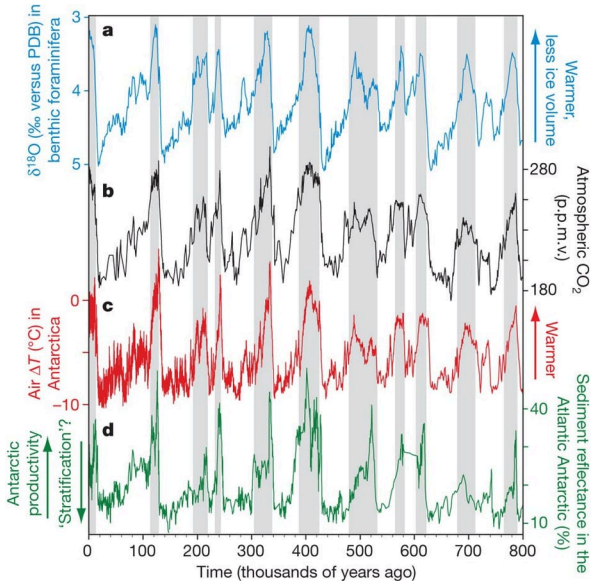



Figure 12.3.8: This graph illustrates the variations in sea levels (blue), atmospheric $CO_2$ concentrations from ice cores (black), Antarctic temperature anomalies measured using deuterium ratios measured in the ice (red), and Antarctic productivity (green) as a function of time. (CC BY-NC 4.0; Daniel M. Sigman et al., 2010 via Nature). Alternative Text Description.

A closer inspection of leads and lags shows that during the last deglaciation, $CO_2$ and Antarctic temperatures lead global temperature and sea levels, suggesting that $CO_2$ is an important forcing mechanism for the warming. But we also know that climate affects the carbon cycle. For example, $CO_2$ is more soluble in colder water, which means the ice-age ocean can take up additional carbon from the atmosphere. Another likely reason for the lower atmospheric glacial $CO_2$ concentrations is iron fertilization. The colder glacial atmosphere was also dustier. Dust contains iron, and thus, iron deposition in iron-depleted regions, such as the Southern Ocean, increased during the ice age. This intensified phytoplankton growth and the biological pump, which are together called Productivity (Figure 12.3.8). This increased phytoplankton activity in the Southern Ocean intensifies sinking and carbon sequestration in the deep ocean. Recent research indicates that about half (~45 ppm) of the glacial-interglacial $CO_2$ variations can be explained by temperature, and another 25-35 ppm by ocean iron fertilization. Thus, we can conclude that climate and the carbon cycle are tightly linked.

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# 12.4: Causes of Climate Change

## Processes that affect Climate

Many processes can alter the climate. These include changes in the Sun's energy output over years; the positions of the continents over millions of years; the tilt of Earth's axis; the orbit over thousands of years; and sudden, dramatic changes due to random catastrophic events, such as a significant asteroid impact.

## Plate Tectonics

Plate tectonics is the scientific theory that Earth's lithosphere comprises several large tectonic plates, which have been slowly moving since 3–4 billion years ago, as illustrated in Figure 12.4.1. Such plate movement and location can alter the climate. Over millions to tens of millions of years, as seas open and close, ocean currents may distribute heat differently. For example, when all the continents are joined into one supercontinent (such as Pangaea), nearly all locations experience a continental climate, except for a small coastal area. When the continents separate, heat is more evenly distributed because there's more coastal land, and even interior land is, on average, closer to the ocean.

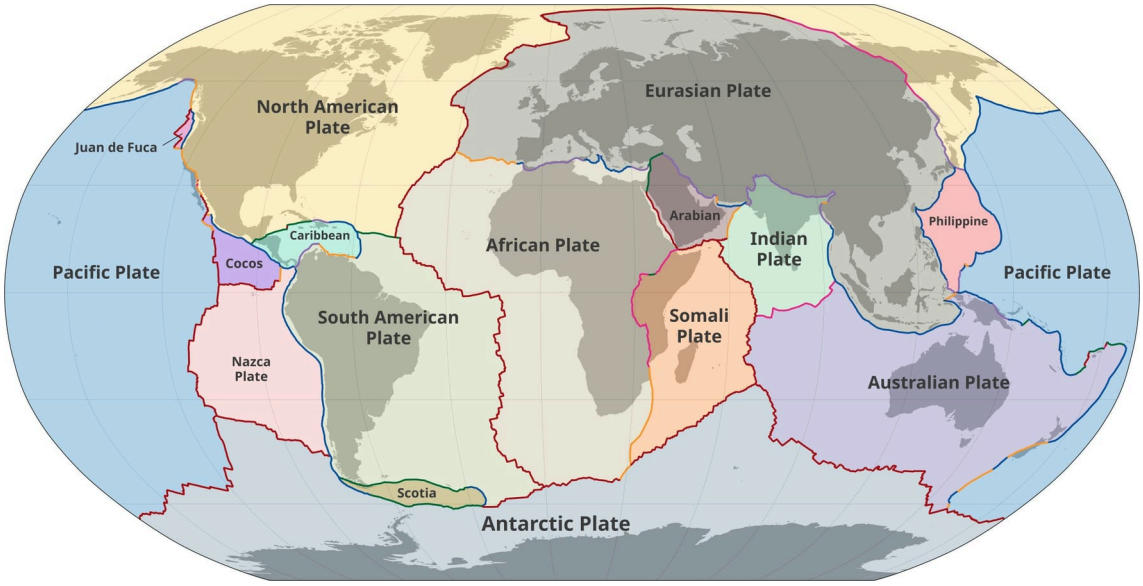



Figure 12.4.1: A map of Earth's 16 principal tectonic plates, which are constantly moving, subducting, and converging as they float over the Earth's mantle. (CC BY-SA 3.0; M. Bitton via Wikimedia Commons). Alternative Text Description.

Plate tectonic movements may also help start an ice age. When continents are located near the poles, ice can accumulate, thereby increasing albedo and lowering global temperature. Low enough temperatures may start a global ice age. Plate motions can also trigger volcanic eruptions, which release dust and $CO_2$ into the atmosphere. Ordinary eruptions, even large ones, have only a short-term effect on the weather. Massive eruptions that create lava plateaus release much more gas and dust and can change the climate for many years. This type of eruption is exceedingly rare; none has occurred since humans have lived on Earth.

## Solar variations

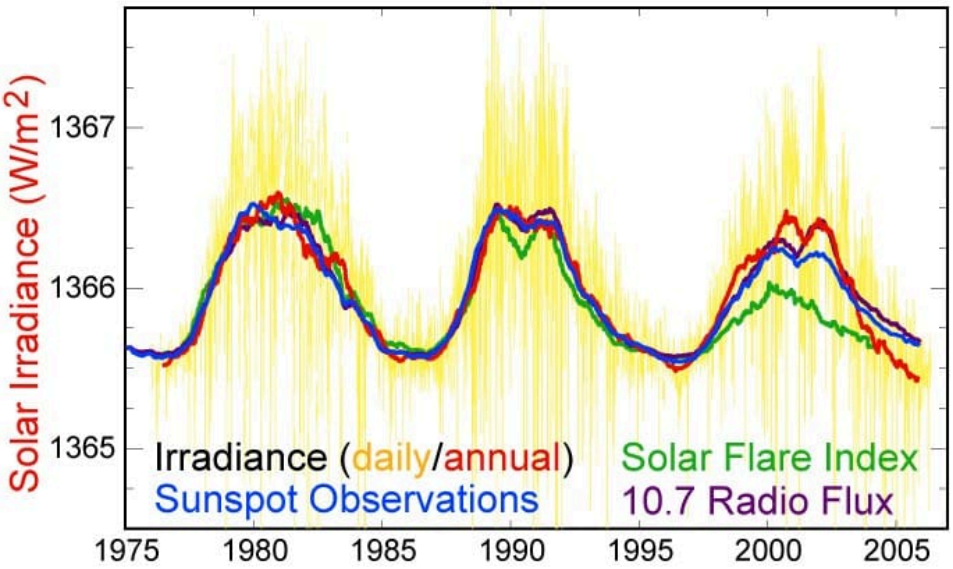



Figure 12.4.2: The last three solar cycles as measured in solar irradiance, sunspot numbers, solar flare activity, and 10.7 cm radio flux. Solar irradiance, i.e., the direct solar power at the top of Earth's atmosphere, is presented as both a daily measurement and a moving annual average. (CC BY-SA 3.0; Robert A. Rohde via Wikimedia Commons). Alternative Text Description.

Stars have life cycles. Some grow, some contract. The Sun continues to grow, and over the last 4.6 B years, its size and energy output have increased by approximately 33%. The amount of energy the Sun radiates is variable, as shown in Figure 12.4.2 . **Sunspots** are magnetic storms on the Sun's surface that increase and decrease over an 11-year cycle. When the number of sunspots is high, solar radiation is also relatively high. The Little Ice Age coincided with a period of solar inactivity, known as the **Maunder Minimum**. But the entire variation in solar radiation is tiny relative to the total amount, and there is no known 11-year cycle in climate variability. We do not see the same 11-year cycle in global temperatures. Instead, we see a steady increase in temperatures over the last century.

## The Milankovitch Cycles

The most extreme climate in Earth's recent history was the Pleistocene. This time period began 2 million years ago. Scientists attribute a series of Pleistocene ice ages to variations in Earth's orbit and tilt relative to the Sun, known as **Milankovitch cycles**. They are caused by changes in Earth's orbit around the sun, which affects the seasonal distribution of incoming solar radiation. This theory was first proposed in 1938 by Serbian astronomer Milutin Milankovich, who calculated variations in Earth's orbital parameters and linked those to past ice ages. Earth's orbit can be described by three parameters, as illustrated in Fig. 12.4.3. The eccentricity $E$ is the deviation from a perfectly circular orbit. Earth's orbit is a slight ellipse, although it is close to circular. $E$ varies on ~100,000-year cycles between zero and 0.06. The tilt $T$, or obliquity, describes the angle between Earth's axis of rotation and the ecliptic, which is the plane of Earth's orbit around the sun. It is currently $T$ = 23.5°, but it varies between 22.5° and 24.5° on a 40,000-year cycle. Precession $P$ is the wobble of Earth's axis, like the wobble of a top. Currently, we're closest to the sun in January, which corresponds to the axis tilted towards the left. $P$ varies on a 23,000-year cycle and is strongly modulated by eccentricity. These variations are caused by gravitational forces from the other planets, particularly Jupiter and Saturn.

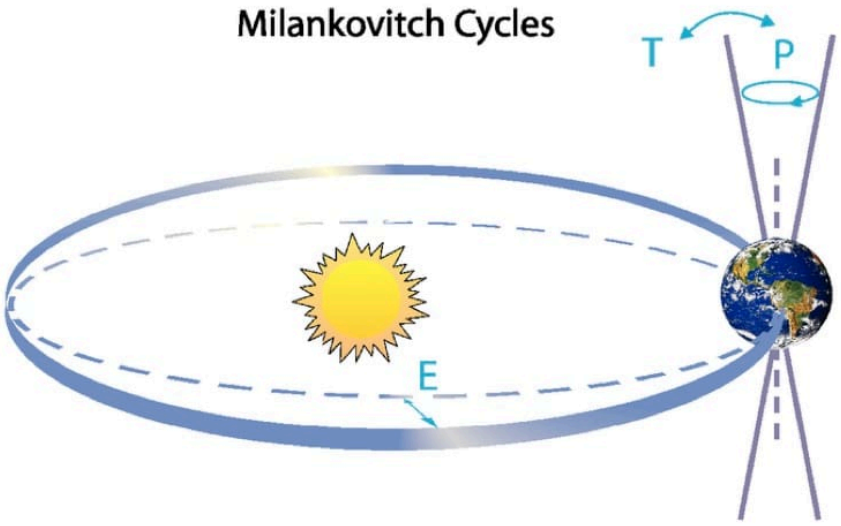



Figure 12.4.3: The Earth moves around the Sun in an elliptical orbit (eccentricity, E), so the distance to the Sun, and therefore the amount of solar energy reaching the Earth, varies periodically. Furthermore, since the Earth's equatorial plane is tilted relative to its orbital plane (obliquity or tilt, T), the amount of energy received from the Sun varies with the seasons. Third, the Earth's axis of rotation spins around the normal to its orbital plane (precession, P). (CC BY-NC 4.0; Mathematics of Planet Earth via Rutgers University). Alternative Text Description.

**Milankovitch's theory** suggests that summer insolation at 65°N controls the waxing and waning of the ice sheets, as illustrated in Figure 12.4.4. When summer isolation is high, all the snow from the previous winter will melt away. When summer insolation is low, some snow survives, and during the following winter, more snow accumulates. In this way, an ice sheet can grow. The Northern Hemisphere is important because it contains the major landmasses and is where additional ice sheets can grow. The Antarctic ice sheet did not fundamentally change during the ice ages, although it grew somewhat larger during them and shrank a bit during interglacials. The Patagonian ice sheet was much smaller than those in the north.

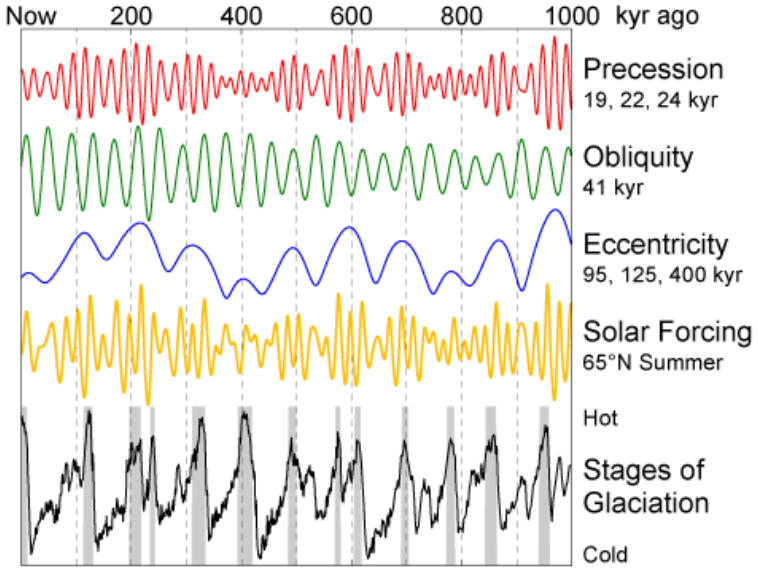



Figure 12.4.4: Variation of Earth's orbital parameters align with Solar Insolation, and the glacial-interglacial cycles. The interplay of the three orbital cycles affects the amount of solar radiation received at different latitudes over the year. The amount of solar radiation reaching the Northern Hemisphere at 65°N seems to control the advance and retreat of glaciers and ice sheets. (Public Domain; Causes of Climate Change via NASA). Alternative Text Description.

While the Milankovich theory explains the cyclicity and timing of glacial-interglacial cycles, it does not explain their amplitudes (how much global average temperatures have changed). Simulations with global climate models show that the amplitude of glacial-interglacial temperature changes can be reproduced only when $CO_2$ changes are accounted for. This leads us to conclude that $CO_2$ changes are an important (feedback) factor in determining glacial-interglacial temperature changes. However, the

ultimate cause of ice-age cycles is Earth's orbital cycles.

## Volcanoes

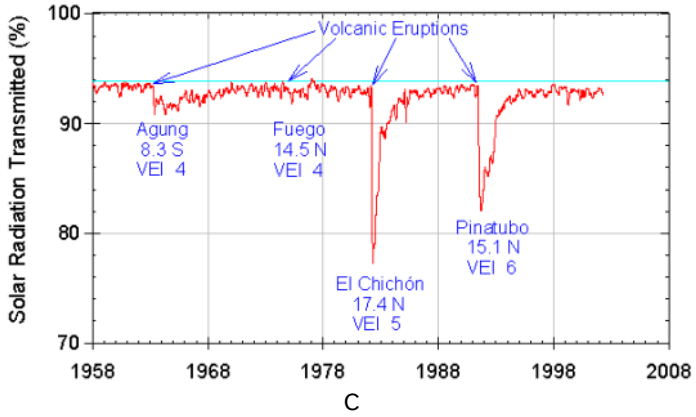
Aerosols are small particles suspended in the air. Natural processes that deliver aerosols into the atmosphere are dust storms and volcanic eruptions. The combustion of oil and coal also releases aerosols into the atmosphere. Aerosols have two main effects on Earth's radiative balance. They directly reflect sunlight back to space (direct effect). They also act as cloud condensation nuclei, forming more or brighter clouds that reflect more solar radiation back to space (an indirect effect). Thus, both direct and indirect effects of increased aerosol concentrations lead to surface cooling. Therefore, aerosol forcing is negative.



A



B




C

Figure 12.4.5: A: Photograph of a rising Mt. Pinatubo ash plume on June 12, 1991 (Public Domain; Dave Harlow, USGS via Wikimedia Commons). B: A photograph from the Space Shuttle over South America taken on Aug. 8, 1991, showing the dispersion of the aerosols from the MT. Pinatubo eruption in two layers in the stratosphere above the top of cumulonimbus clouds (Public Domain; Stephen Self et al., 1993 via USGS). C: Measurements of solar radiation transmitted at Hawaii's Mauna Loa Observatory show a reduction in solar incoming radiation during major volcanic eruptions. (Public Domain; Kokoo via Wikimedia Commons). Alternative Text Description.

On June 12, 1991, Mt. Pinatubo in the Philippines erupted spectacularly, sending an ash column 19 km (12 mi) into the atmosphere, as shown in Figure 12.4.5A. Large explosive volcanic eruptions such as these can eject ash particles and gases such as sulfur dioxide ($SO_2$) into the stratosphere (Figure 12.4.5B), where they can be rapidly distributed over large areas. In the stratosphere, $SO_2$ is oxidized to form sulfuric acid aerosols. However, stratospheric aerosols eventually mix back into the troposphere and are removed through precipitation or dry deposition. The lifetime of volcanic aerosols in the stratosphere is in the order of months to a few years. Estimates of the radiative forcing from volcanic eruptions depend on the eruption but vary from a few negative tenths of a watt per square meter to -3 or -4 $Wm^{-2}$ for the most significant eruptions during the last 100 years, as illustrated in Figure 12.4.5C.

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# 12.5: Recent Climate Change

## Changes Over the Past 100 Years

Although Earth's climate is currently changing rapidly relative to past changes, in most regions where we live, changes are slow enough that we do not notice them directly during our daily lives. However, older people may have observed changes over their lifetimes, and in some regions, these changes are larger and more pronounced than elsewhere. In this section, we will discuss observations from the past 100 years and data from regions susceptible to climate change, where the most dramatic effects have occurred. This will not be a comprehensive documentation of existing observations. Additional observations will be discussed throughout the remainder of this book.

## The Climate System

Earth's climate system consists of interacting **components**, as depicted in Fig. 12.5.1. The **atmosphere**, which is the air and clouds above the surface, is about 10 km thick (more than two-thirds of its mass is contained below that height). The **ocean** covers more than two-thirds of Earth's surface and has an average depth of roughly 4 km. Contrast those numbers with Earth's radius, which is approximately 6,400 km, and you'll find that Earth's atmosphere and ocean are very thin layers compared to the size of the planet itself. In fact, they are about 1,000 times thinner. They are perhaps comparable to the outer layer of an onion or to the water on a wet soccer ball. Yet all life is constrained to these thin layers. The major ocean basins are the Pacific, Atlantic, Indian, and Southern Oceans. The **Cryosphere** comprises ice and snow, including sea ice, mountain glaciers, and ice sheets. Sea ice is frozen seawater, up to several meters thick, floating on the ocean. Ice sheets on land, composed of compacted snow, can be several kilometers thick. The **Biosphere** includes all living things on land and in the sea, from the smallest microbes to trees and whales. The **Lithosphere**, which is the solid Earth (upper crust and mantle), could also be considered an active part of Earth's climate system because it responds to ice load_and influences atmospheric carbon dioxide ($CO_2$) concentrations and climate on long timescales through continental movements.



Figure 12.5.1: A composite image of Earth from space. It shows all four components of Earth's climate system. The atmosphere has its complex cloud patterns. The ocean covers about 70% of Earth's surface. The cryosphere is visible as the white areas on the top: sea ice covering the Arctic Ocean and the Greenland ice sheet. Green colors on land and turquoise shades along the ocean margin indicate the biosphere, including forests and phytoplankton blooms. Notice the thin layer of the atmosphere surrounding Earth in the lower left. (Public Domain; The blue marble via NASA)

The components interact by exchanging energy, water, momentum, and carbon, creating a deliciously complex coupled system. Imagine water evaporating from the tropical ocean heated by the sun. The air containing that water rises and cools, which causes the water vapor to *condense* into a cloud. The cloud is carried by winds over land where it rains. The rain sustains a forest. Trees

are dark, having a low albedo. This influences the amount of sunlight absorbed by the Earth. Dark surfaces absorb more sunlight and warm up more than bright surfaces, such as desert sand or snow. Air warmed by the surface rises, affecting the wind.

## Atmosphere

Observations indicate that the global climate is changing. Surface air temperature data, averaged over the whole Earth, indicate warming of 0.7-0.9°C over the last ~100 years, as shown in Figure 12.5.2. But that warming was neither steady nor smooth. From 1880 to about 1910, there was cooling, followed by warming until about 1940. After this slight cooling, approximately constant temperatures were observed until the 1970s, after which temperatures rose rapidly to the present. Each year's temperature differs slightly from the next. Not all of these year-to-year changes are currently understood, but natural variations, which humans do not cause, do play a role. For instance, strong El Niño years, such as 1997-98 or 2015, are particularly warm, whereas La Niña years, such as 1999-2000 or 2011, are relatively cool. El Niño and La Niña are also known together as ENSO (El Niño/Southern Oscillation). ENSO is a climate variability pattern occurring in the tropical Pacific that also impacts many other regions of the Earth.

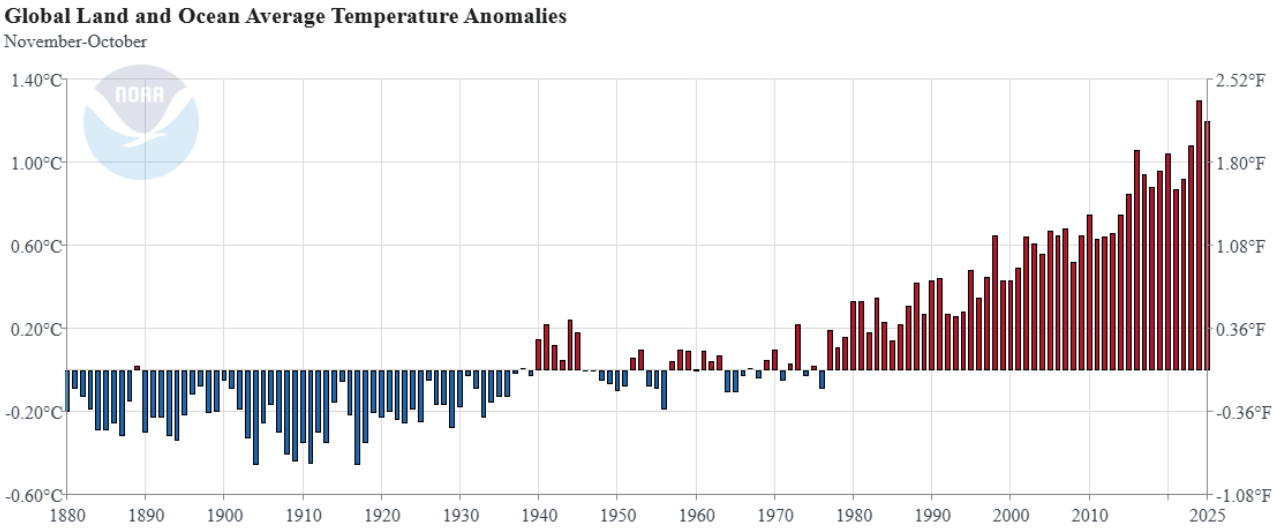



Figure 12.5.2: Yearly surface temperature from 1880–2025 compared to the 20th-century average (1901-2000). Blue bars indicate cooler-than-average years; red bars show warmer-than-average years. (Public Domain; Global Time Series via NOAA/NCEI)

The increase in atmospheric temperatures over the last 100 years has not been uniform worldwide. Figure 12.5.3 shows that temperatures over land changed more than over the oceans, and the Arctic warmed more than the tropics. These patterns, called *land-sea contrast* and *polar amplification,* are pretty well understood and simulated in climate models, as we will see later. The warming is indeed almost global. The only exception is the northern North Atlantic, which has been slightly cooling for reasons we will discuss later.

Global surface temperature change map shows high temperature increases in the polar regions.

Figure 12.5.3: Global map of recent surface temperature change shows that the warming is not uniform. Higher latitudes are warming faster than the tropics, while some oceanic regions are cooling. (CC BY-SA 4.0; NASA via Wikimedia Commons)

Analysis of thousands of temperature records into an estimate of global mean temperature change, such as Fig. 12.5.2, is not trivial. Issues such as changes in station locations, instrumentation, and data coverage have to be taken into account. Thus, there are a few other groups around the world that calculate global average surface temperatures, e.g. the National Aeronautics and Space Administration's (NASA) Goddard Institute for Space Science (GISS), the Climate Research Unit (CRU) of the University of East Anglia, or the Berkeley Earth Surface Temperatures (BEST), and all come to very similar results. Errors due to unequal sampling have been estimated at approximately 0.2°C during the early part of the record and decreasing to 0.1°C during the more recent period. The fact that five groups analyzing the data with different methods reach the same conclusions suggests that the results are robust. The data's reliability has been demonstrated, and station locations (e.g., urban versus rural) don't matter.

The global warming trend has increased the likelihood of warm and extremely warm temperatures and decreased the likelihood

of cold and extremely cold temperatures. E.g., extremely warm (more than 3°C above the 1951 to 1980 average) summer temperatures over Northern Hemisphere land areas had only a 0.1% chance of occurring from 1951 to 1980. In contrast, during the decade from 2005 to 2015, the chance of those extreme warm temperatures occurring was 14.5%. Conversely, relatively cold summer temperatures (about 0.5°C cooler than the 1951-1980 average) that occurred about every third year from 1951 to 1980 occurred only 5% of the time during 2005 to 2015.

## Cryosphere

One of the most sensitive regions to climate change is the Arctic. Sea ice cover there has decreased dramatically over the past 40 years, particularly in late summer (Figure 12.5.4). Arctic sea ice experiences a strong seasonal cycle. In late winter, it covers about $15.4 \times 10^6$ km$^2$ and decreases to about $6.4 \times 10^6$ km$^2$ in late summer, as depicted in Figure 12.5.5. Therefore, the relative changes are larger in late summer, with a reduction of about 46 % or $2.9 \times 10^6$ km$^2$ from 1980 to 2015. In winter, the reduction was only about 9 % or $1.5 \times 10^6$ km$^2$. The lost area of summer Arctic sea ice is more than 10 times the size of Oregon (255,000 km$^2$) or four times the size of France (~650,000 km$^2$).

Arctic Sea Ice September 1979 shows frozen waters from Siberia to Canada across the north pole.

A

Arctic Sea Ice September 2020 shows no sea ice near Siberia any more.

B

Figure 12.5.4: Arctic sea ice extent in September 1979 (A) and 2020 (B) from satellite observations. The purple line denotes the median ice extent for 1981-2010. In 1979, the ice extent was 7.1 million square kilometers; in 2020, it was 3.9 million square kilometers. (Public Domain; Sea ice extent via NOAA/NSIDC)

In the Antarctic, sea ice undergoes a larger seasonal cycle, but it has changed less over the last 40 years than in the Arctic. Southern hemisphere sea ice has slightly increased by about 12 % or $0.5 \times 10^6$ km$^2$ in late austral summer and 2.4 % or $2.3 \times 10^6$ km$^2$ in austral winter. Note that year-to-year fluctuations are larger in winter than in summer and that the long-term trends in the Arctic are much larger than the short-term fluctuations. In contrast, in the Antarctic, the long-term trends are similar to the short-term fluctuations and thus less statistically significant. A trend is statistically significant if it exceeds the errors.

Graphs showing northern, southern and global sea ice reducing steadily over the last 50 years. Details in Caption.

Figure 12.5.5: Changes in late summer sea ice extent in the Arctic (top) and Antarctic (bottom), and global monthly sea ice extent (center). (Public Domain; Global Snow and Ice Report via NOAA/NSIDC). Alternative description of the image.

Globally, Earth has lost approximately 2 million square kilometers of sea ice from 1980 to 2015, representing about 10% of the total. Compare that area to that of your favorite state or country.

**Mountain glaciers** are also sensitive to climate change. Figure 12.5.6 shows an example from Muir Glacier, which has retreated dramatically since 1941. This is typical of most glaciers worldwide. In fact, only a small number of glaciers show advances, whereas the vast majority melt and retreat from the valleys into higher elevations.



Figure 12.5.6: Muir Glacier in Glacier Bay National Park and Preserve, Alaska. The older photograph (A) is from 1941, and the recent photograph (B) shows the Muir Glacier in 2004. (Public Domain; Glaciers via NOAA/NSIDC).

The World Glacier Monitoring Service (WGMS) has compiled information on hundreds of glaciers worldwide. Figure 12.5.7 shows that since 1980, glaciers in all regions have been losing mass with an acceleration of loss in recent years.

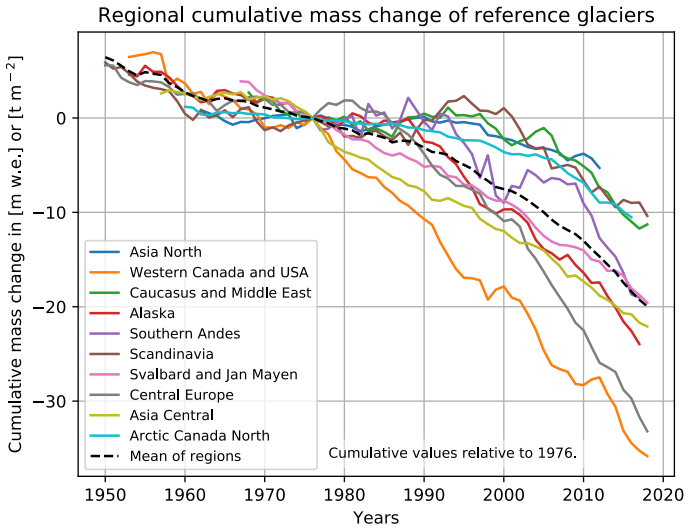



Figure 12.5.7: Cumulative mass balance in meters water equivalent (w.e.) of mountain glaciers decreasing steadily in the past 70 years. (CC BY-NC 4.0; Global Glacier State via World Glacier Monitoring Service). Alternative description of the image.

Ice sheets are massive, continental-scale bodies of ice and snow that cover more than 50,000 square kilometers of land. They form over thousands of years when annual snowfall exceeds melting, causing layers of snow to compact into flowing ice. The two current ice sheets on Earth are in Greenland and Antarctica, which together hold over 99% of the planet's freshwater ice. Observations from the Gravity Recovery and Climate Experiment (GRACE) satellites, which measure Earth's gravity field with very high precision and can detect mass changes, indicate that since 2002 the Greenland ice sheet has lost approximately 4,000 Gt of mass. The Antarctic ice sheet has lost about 2,500 Gigatons as illustrated in Fig. 12.5.8. The melting of the Greenland ice sheet currently contributes approximately 0.8 mm/yr to global sea-level rise. Antarctica's contribution is 0.4 mm/yr, and mountain glaciers add about 0.6 mm/yr, for a total of 1.8 mm/yr sea level rise from melting of ice, according to IPCC 2019 reports.

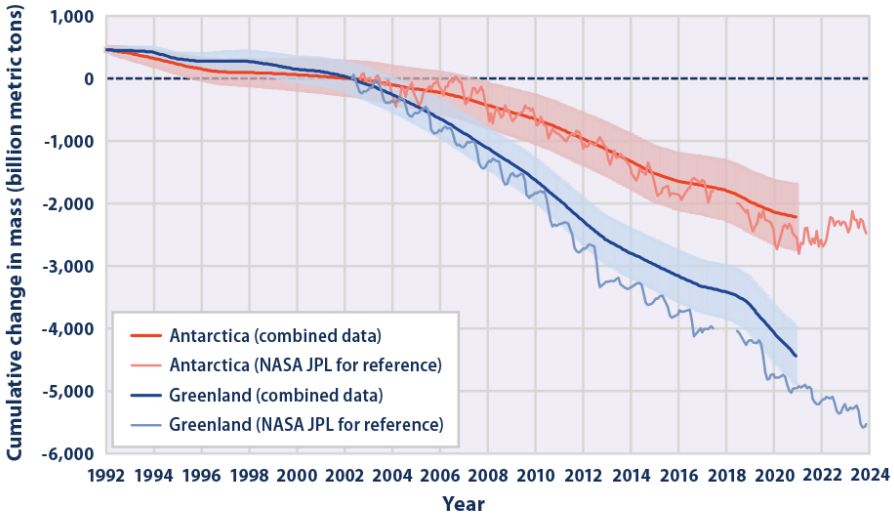



Figure 12.5.8: Changes in Antarctic (red) and Greenland (blue) ice mass in billion metric tons. The dark lines are based on more than 20 studies combined for each region. Shading shows uncertainty estimates from the various data sets that contribute to the combined average. The two "NASA JPL" lines have been added to show results from one commonly cited analysis, including seasonal variations. All estimates are centered at zero in 2002 to provide a consistent point of reference. Thus, a downward slope indicates a net loss of ice and snow. For reference, 1,000 billion metric tons is equal to about 260 cubic miles of ice—enough to raise sea level by about 3 millimeters. (Public Domain; NASA JPL via EPA). Alternative description of the image.

## Oceans

Subsurface ocean temperature measurements document warming over the last 60 years. The ocean's **heat content** has increased by about $30 \times 10^{22}$ Joules during that time, as illustrated in Fig. 12.5.9. Before 2005, subsurface temperature measurements were more limited in space and time because they were taken from ships by lowering CTD (conductivity, temperature, depth) instruments on a cable into the ocean. Since 2005, autonomous, free-drifting Argo floats have measured temperature, salinity, pressure, and velocity of the upper 2 km of the water column. Currently, there are approximately 4,000 floats in operation, providing substantially better spatial and temporal coverage than previous ship-based measurements.

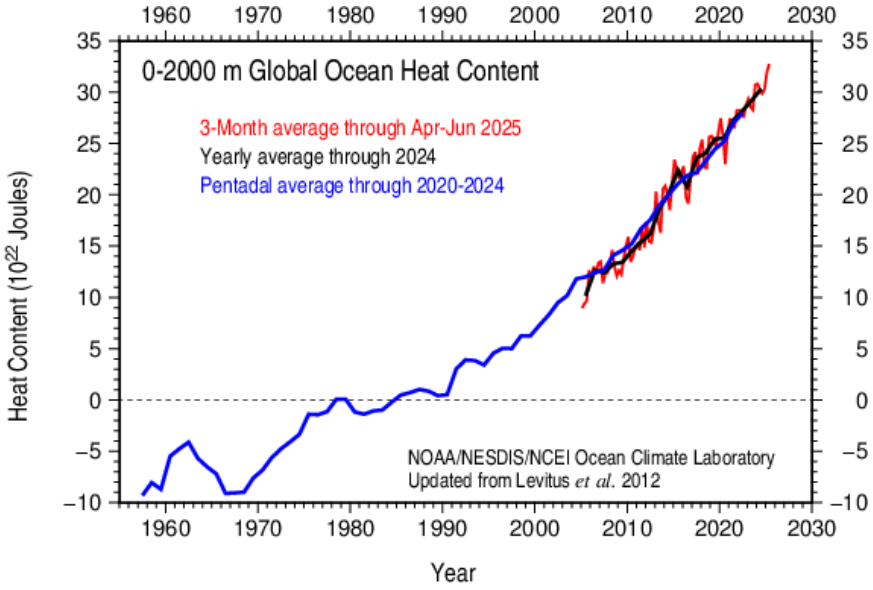



Figure 12.5.9: Observational estimates of global ocean heat content (0-2 km). Between 1960 and 2005, ocean temperatures were measured mainly using research ships (blue line). Since 2005, autonomous ARGO floats have increased data density in both space and time, as shown by the red and black lines. Both sets of measurements indicate that global ocean heat content has increased significantly in the last 60-70 years. (Public Domain; NASA JPL via EPA). Alternative description of the image.

The melting of mountain glaciers and ice sheets leads to increased runoff into the ocean, which contributes to **sea level** rise as shown in Fig. 12.5.10. Sea level rise is also caused by warming sea water, which expands, and by increased runoff from groundwater pumping from aquifers. Estimates based on tide gauge records indicate that sea level has risen by approximately 20 cm from the 1870s to 2000, and by an additional 6 cm since then. The melting of mountain glaciers and ice sheets contributes approximately equally to current sea-level rise; however, if current trends continue, many mountain glaciers will likely disappear, and the large ice sheets will contribute more and more to global sea-level rise. Note that sea level rise is not spatially uniform.

Graph showing a steady increase in global sea levels over the last 140 years.

Figure 12.5.10: Observed global mean sea level estimated from satellites (red) and tide gauges (blue). The pie chart shows the relative contributions from the melting of land ice (1.8±0.1 mm/yr), thermal expansion of seawater (1.4±0.3 mm/yr), and groundwater pumping (0.4 mm/yr; estimated as the residual) during 2005-2015 (IPCC, 2019). The total rate of sea level rise during this period (3.6±0.5 mm/yr) is larger than during earlier periods. This acceleration in sea-level rise is driven by the recent increase in melting of the Greenland and Antarctic ice sheets. (CC BY-NC 4.0; Andreas Schmittner via Introduction to Climate Science). Alternative description of the image.

## Biosphere and Carbon Cycle

Plant and animal species have been observed moving poleward and upward in mountain ranges. This response is consistent with global warming and with organisms' tendency to remain within a temperature range to which they have adapted. Flowering dates of many plants have also shifted earlier in the spring, including cherry blossoms.

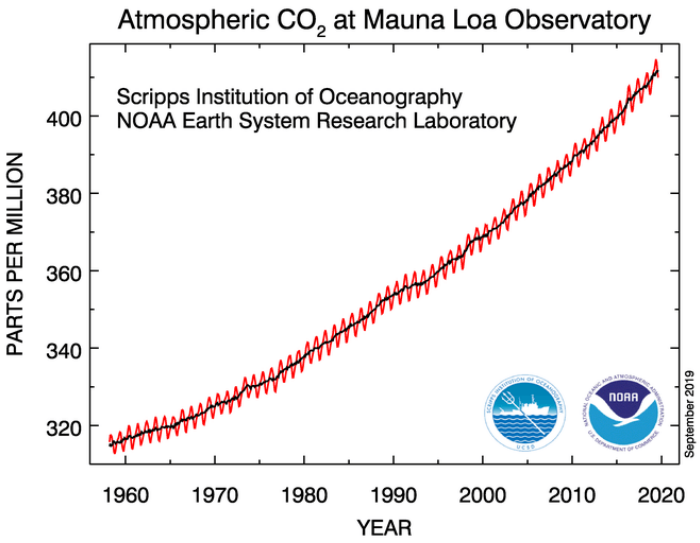



Figure 12.5.11: Atmospheric $CO_2$ measured at Hawaii's Mauna Loa Observatory. The measurements were pioneered by Charles Keeling from Scripps Institution of Oceanography in 1958. (Public Domain; Global Monitoring Laboratory via NOAA). Alternative description of the image.

Carbon dioxide ($CO_2$) has been measured in the atmosphere since 1958 at Mauna Loa Observatory in Hawaii, as illustrated in Figure 12.5.11. At that time, concentrations were just below 320 parts per million (ppm). Subsequently, they increased to values of just over 424 ppm today. That is a 25% increase. Overlaid on the long-term trend is a seasonal cycle. Growth of the terrestrial biosphere in the Northern Hemisphere spring leads to CO2 drawdown and the decay of organic matter, such as fallen leaves, which increases $CO_2$ in the fall. $CO_2$ is an important greenhouse gas, and its increase over the past decades is the leading cause of the recent global warming.

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# 12.6: Solutions

## Global Emissions

Climate change is a complex problem because modern society was built around burning fossil fuels for energy. We still depend heavily on fossil fuels for everything from driving our cars and washing our laundry to charging our cell phones and heating our homes. To stabilize the climate, however, we'll need to reach near-zero carbon emissions over the long term. Thus, the challenge is to decarbonize our economy. The longer we wait, the greater the impacts of future climate change will be, or the faster we'll need to reduce emissions. And because it is a global problem, the whole world, or at least most of it, will need to cooperate to solve it. Moreover, because we're already committed to further climate change, we'd better prepare to adapt. This chapter discusses technological options, political efforts, and personal actions for climate mitigation and adaptation to the inevitable.

Figure 12.6.1 illustrates the increase in carbon dioxide ($CO_2$) emissions since 1750. Before the Industrial Revolution, emissions were very low. Emissions grew relatively slowly until the mid-20th century. In 1950, the world emitted 6 billion tonnes of $CO_2$. By 1990, this had nearly quadrupled to more than 20 billion tonnes. Emissions have continued to grow rapidly. We now emit more than 38 billion tonnes annually. Emissions growth has slowed in recent years, but they have not yet peaked. In recent years, China, the United States, and India have been the top three $CO_2$ emitters by total volume, with China being the largest emitter since 2006. Other major emitters include Russia, European Union (EU) member states, and Indonesia. While emissions from China and India have increased over the past few decades, their per capita emissions remain significantly lower than those of countries such as the United States and the EU.

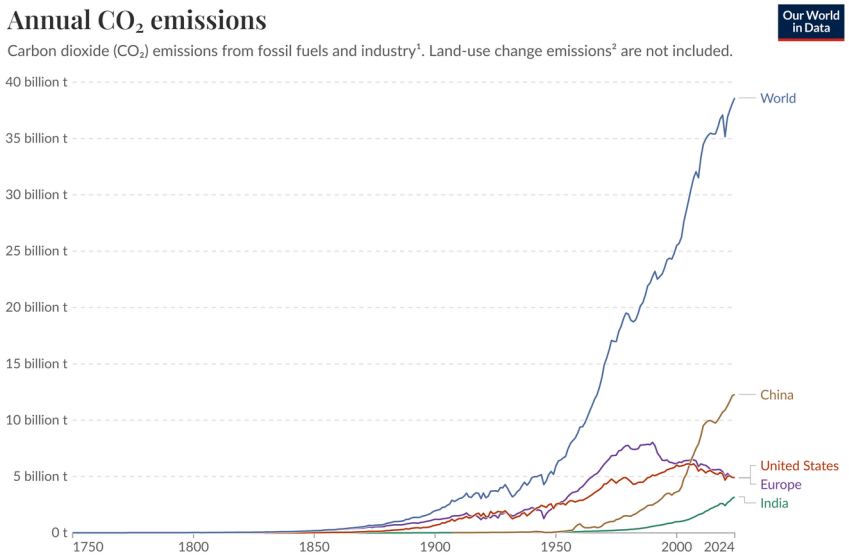



Figure 12.6.1: The growth of global emissions from the mid-18th century through to today. Growth in emissions was still relatively slow until the mid-20th century. In 1950, the world emitted 6 billion tonnes of $CO_2$; by 2024, it had risen to over 38 billion tonnes. (CC BY 4.0; Global Carbon Budget (2025) via Our World in Data.)

## Technology

Current global energy production relies heavily on fossil fuel burning, as illustrated in Fig. 12.6.2. The largest energy sources are oil, coal, and natural gas, all fossil fuels, while non-fossil fuels together account for only about 20% of total energy. In the United States, transportation is the largest source of emissions, accounting for 27%, followed by electricity generation at 25% and industry at 24%.

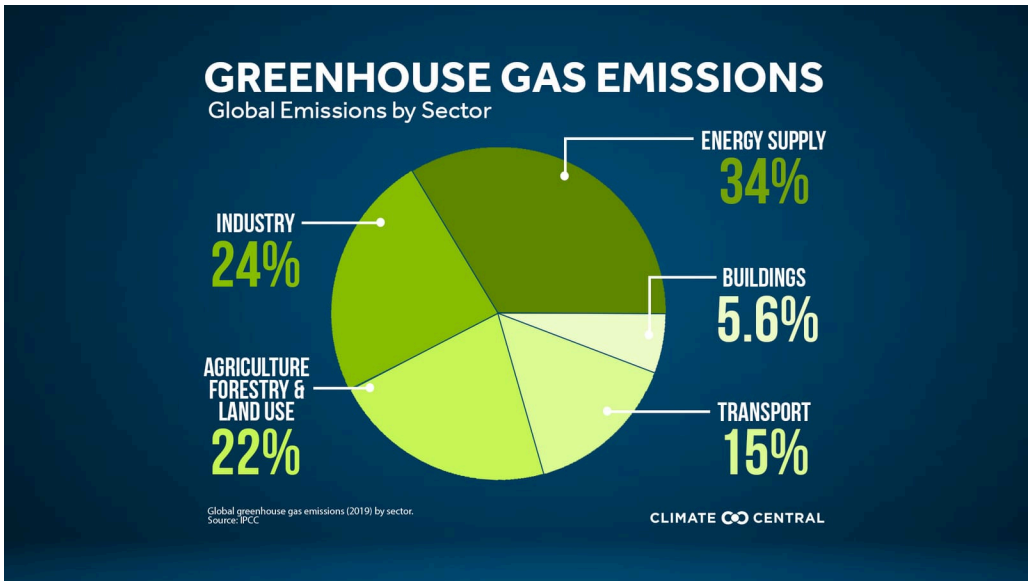



Figure 12.6.2: This pie chart illustrates the breakdown of global greenhouse gas emissions by sector in 2019. The energy supply sector is the most significant contributor at 34%, followed by industry (24%), agriculture (22%, which includes forestry and land use), transportation (15%), and buildings (5.6%). (CC BY-NC 4.0; Peak $CO_2$ and heat trapping emissions via Climate Central)

Most oil consumption powers transportation via internal combustion engines in cars and trucks, which are very inefficient. Only about 25% of the energy input into transportation moves vehicles, while nearly 75% is wasted as heat. Coal is mainly used in power plants to generate electricity, and this process also loses about 50% of the total energy. This loss is less than that of internal combustion engines, so electric cars have lower carbon footprints than internal combustion engine cars, even if the electricity comes from coal. Most natural gas is used to generate electricity and to heat buildings. Hydro- and nuclear power generate electricity, whereas biomass is primarily used for cooking and home heating in the developing world. Renewable energy sources, such as solar and wind, supply only a small fraction of total energy.

## Nuclear Power

Nuclear power generates electricity by using heat from a controlled nuclear fission reaction to create steam, which drives a turbine connected to a generator. They provide fossil-free power that does not emit carbon dioxide (except during construction). Unlike intermittent sources such as solar and wind, nuclear power plants can operate 24/7, providing a steady, reliable baseload power supply. For this reason, some view them as an important future energy source. The Diablo Canyon Nuclear Power Plant depicted in Figure 12.6.3 is the only operational nuclear plant in California, as well as the state's largest single power station.



Figure 12.6.3: The Diablo Canyon Power Plant near Avila Beach in San Luis Obispo County in California. (CC BY 2.0; Tracy Adams via Flickr)

However, nuclear power has disadvantages too. They are expensive to build, dangerous to operate, and produce radioactive waste for which no long-term repository currently exists. Catastrophic accidents such as the nuclear meltdowns in 2011 at the Japanese Fukushima Daiichi plant and in 1986 at the Chernobyl reactor in what is now northern Ukraine have shown the dangers associated with nuclear power production. As a result, nuclear power remains controversial.

## Renewables

In some countries, renewable energy sources have increased rapidly in recent years. Germany, for example, increased the share of renewables in total electricity production from 3% in 1990 to 45% in 2020, while its economy has remained among Europe's strongest. Denmark plans to move to 100% renewable energy by 2050. As of 2023, in the United States, renewables currently account for 9% of total energy consumption and 21% of electricity production, as illustrated in Figure 12.6.4. Because transportation consumes the most energy in the United States, petroleum (oil) accounts for most of that energy. Fossil fuels such as natural gas and coal still account for nearly 60% of electricity generation, with nuclear power and renewables accounting for the remainder.

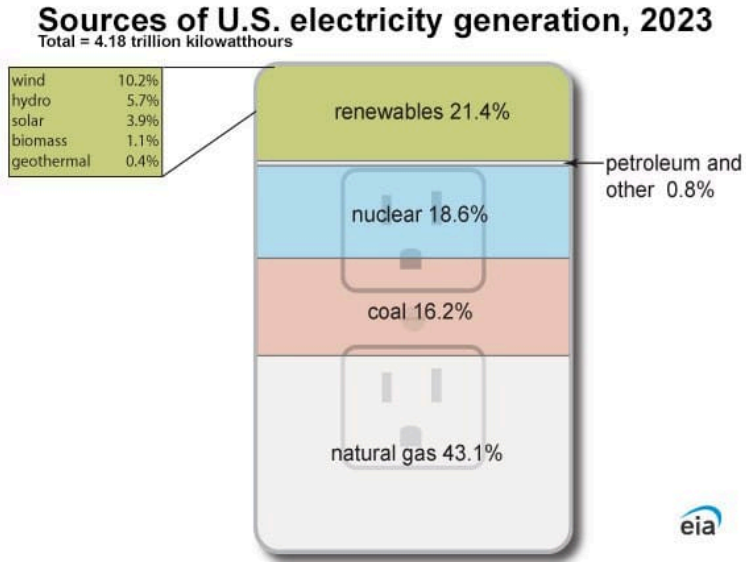
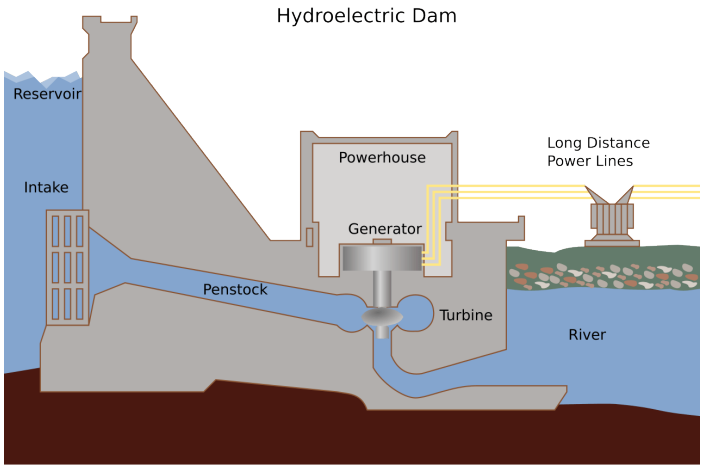



Figure 12.6.4: Electricity generation in the United States comes primarily from natural gas (43.1%) and coal (16.2%). Nuclear power accounts for 18.6%, and renewable sources, including geothermal, solar, hydroelectric, wind, and biomass, account for only 21.4%. (Public Domain; Electricity Explained via U.S. EIA). Alternative Text Description.

**Solar power** is the conversion of sunlight into electricity, either directly via photovoltaic (PV) cells (also known as solar panels) or indirectly via concentrated solar power using mirrors. **Wind power** uses wind energy to rotate wind turbines, which are then connected to a generator to produce electricity. A renewable energy farm including solar panels and wind turbines in Germany is shown in Figure 12.6.5A. **Hydro power** is generated by converting the gravitational potential or kinetic energy of a water source into electricity, as illustrated in Figure 12.6.5B. **Bioenergy** is a renewable energy source derived from plants and animal waste. Unlike fossil fuels, biomass commonly used for bioenergy includes wood, food crops such as corn, energy crops, and waste from forests, yards, or farms. **Geothermal power** is electricity generated from Earth's core. Water or another fluid is sent down multiple miles below the surface, where the temperature increases. This boils the liquid in the pipes, which then drives a steam turbine at the surface.



A




B

Figure 12.6.5: A: Photovoltaic array and wind turbines at the Schneebergerhof wind farm in the German state of Rheinland-Pfalz. (CC BY-SA 3.0; Kuebi via Wikimedia Commons). B: A schematic diagram of a hydroelectric power station. (CC BY 2.5; Tomia via Wikimedia Commons).
Alternative Text Description.

The advantage of renewable energy is its nearly unlimited supply potential and minimal carbon emissions. Some emissions occur during the production and installation of solar panels and wind turbines. Renewables are also not associated with the dangers of nuclear power. Their main disadvantage used to be their high up-front investment costs. Once installed, however, solar panels and wind turbines require minimal maintenance because solar energy and wind are free. During the past 10 years, the cost of solar panels has decreased dramatically by 80%. Thus, over a system's lifetime, renewables become competitive with fossil fuels. Other renewable energy sources include geothermal, tidal, and wave energy, as well as hydropower. Another issue with renewable energy is its intermittent supply. Solar panels operate during daylight, whereas wind turbines operate only when the wind blows. However, recent studies, such as Shaner et al. (2018) have found that 80% of total electricity demand could be met by wind and solar, as quoted from the abstract of the paper:

> Assuming minimal excess generation, lossless transmission, and no other generation sources, the analysis indicates that wind-heavy or solar-heavy U.S.-scale power generation portfolios could in principle provide ~80% of recent total annual U.S. electricity demand. -Shaner et al. (2018)

Many companies are moving toward **electric** or hybrid vehicles. Because they are much more efficient (80-90%), their energy use is much lower than for internal combustion engine vehicles, and their carbon footprint can be close to zero (some is associated with vehicle production) if they use renewable electricity. Even when electricity is generated from fossil fuels, the carbon footprint of electric cars remains lower than that of internal-combustion-engine vehicles. Given the transportation sector's substantial carbon emissions and significant losses, transitioning to an electric vehicle fleet could substantially reduce future emissions. Although electric cars are more expensive to purchase, their total life-cycle costs are lower than those of gasoline-powered vehicles. Electric cars also offer additional advantages—no oil changes, no tailpipe emissions, and greater torque. However, manufacturing electric cars is not without environmental and human impacts, including those associated with mining the raw materials used in battery production.

## Personal Actions

Each of us can take individual actions to reduce carbon emissions and the impacts of climate change. Conserving energy, e.g. by switching off lights when they are not needed. Switching to energy-efficient appliances, such as LED lights. Recycling also helps. Driving less or choosing an energy-efficient car, such as a hybrid, electric, or smaller vehicle, will reduce your carbon footprint. Even better is riding your bike. Eating less meat is another way to reduce your carbon footprint, since meat production generates more carbon emissions than vegetarian food. Homeowners can weatherize their homes or install solar panels. You can assess your carbon footprint on various websites to help you get started.

Ultimately, however, individual actions have limited effects. For example, imagine you start using less gas by riding your bike instead of driving your car. If many people did, demand for gas would decrease, lowering its price, assuming supply remains constant. Lower gas prices lead people to buy larger cars and drive more, thereby diminishing or erasing the effects of their personal emission reduction. That is not to say that we shouldn't take individual actions. However, collective action is also needed. Large-scale government actions, such as a carbon tax or a cap-and-trade system, can accelerate the transition to renewable energy sources that are necessary to avoid potentially dangerous climate change.

## Adaptation

Climate change already affects us. It is not only a future problem or a faraway issue. It is a present issue. Since we cannot avoid further climate change in the immediate future, we must adapt to avoid its consequences. Perhaps the most important thing is to recognize that it is happening. This will prompt us to consider the potential implications for our region. If regional projections are available, we can consider them in planning. If not, we can still extrapolate from past changes, assuming we'll see more of what we've observed recently.



Figure 12.6.6: The Isle of Wight sea wall, in Ventnor, UK. (Public Domain; Oikos-team via Wikimedia Commons)

Adaptation can help manage impacts and risks to people and nature. The four types of adaptation actions are infrastructural, institutional, behavioral, and nature-based options. Some examples are building seawalls to combat rising sea levels, as shown in Figure 12.6.6, or inland flood defenses. Other options include providing new insurance schemes for populations threatened by climate-related disasters, adjusting crop planting times or varieties based on temperature and precipitation changes, and installing green roofs or green spaces in cities to combat the Urban Heat Island effect. Adaptation can be reactive, i.e., responding to climate impacts as they happen, or proactive, i.e., taking steps in anticipation of future climate change.

## Cited Source:

Shaner, M. R., S. J. Davis, N. S. Lewis, and K. Caldeira, 2018: Geophysical constraints on the reliability of solar and wind power in the United States. *Energy Environ. Sci.*, **11**, 914–925, https://doi.org/10.1039/C7EE03029K.

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# 12.7: Exercises

## Chapter 12 Questions to consider

1. What are the components of Earth's climate system?
2. List processes that cause interactions between the components: atmosphere-ocean, atmosphere-biosphere, atmosphere-cryosphere, ocean-biosphere, ocean-cryosphere.
3. Name three processes (*forcings*) that can cause the global climate to change?
4. What are paleoclimate proxies?
5. Which paleoclimate record has higher resolution: record A, which has data every 100 years, or record B, which has data every 1,000 years?
6. What is the Milankovitch theory?
7. How do we know the extent of glaciation during the last ice age?
8. By how much did global surface air temperatures increase during the last 100 years?
    a. How do we know?
    b. What was the average rate of change during that time?
    c. What was the rate of change for the last 50 years?
    d. Where was the warming larger over the oceans or over land, in the tropics or at high latitudes?
9. What are the two main reasons for the increase in historical carbon emissions?
10. List five things that can be done to mitigate climate change.

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# 12.8: Accessible Descriptions

## Figure 12.1.1: Koppen Climate Classification System

World map showing global climate classifications using the Köppen system. Different regions are grouped into major climate types based on temperature and precipitation patterns. Tropical climates occur near the equator (approximately 0–15 degrees latitude), including rainforest regions in the Amazon, central Africa, and Southeast Asia, characterized by warm temperatures year-round (typically above 18°C) and high precipitation. Surrounding these are monsoon and savanna climates extending to about 10–25 degrees latitude, with seasonal rainfall patterns.

Dry climates dominate subtropical regions (approximately 15–35 degrees latitude), including deserts and semi-arid areas such as the Sahara, Arabian Peninsula, central Australia, and parts of the southwestern United States. These regions receive very low annual precipitation, often less than 250 mm for deserts and up to about 500 mm for steppe climates.

Temperate climates are found in mid-latitudes (about 30–50 degrees), including much of North America, Europe, East Asia, and parts of South America and Australia. These regions have moderate temperatures, with warm summers (often 10–25°C) and mild to cool winters. Subtypes include Mediterranean climates with dry summers, humid subtropical climates with hot summers and year-round precipitation, and marine west coast climates with mild, wet conditions.

Continental climates occur in higher mid-latitudes (about 40–70 degrees), especially in the interiors of North America and Eurasia. These regions experience large seasonal temperature ranges, with warm summers (above 10°C) and cold winters (often below –10°C).

Polar climates dominate regions above about 70 degrees latitude, including Antarctica and Greenland, where temperatures remain below 10°C year-round. Tundra regions have short, cool summers, while ice cap regions remain below freezing throughout the year.

Highland climates appear in mountainous regions such as the Andes, Rockies, Himalayas, and East African Highlands, where climate varies with elevation rather than latitude.

*Table 12.8.1 Detailed data table (Köppen climate classifications):*

| Climate group | Code examples | Temperature characteristics | Precipitation characteristics | Typical regions |
|---|---|---|---|---|
| Tropical | Af, Am, Aw | All months > 18°C | High rainfall (often >1500 mm/year) | Amazon, Congo Basin, Southeast Asia |
| Dry (arid/semi-arid) | BWh, BWk, BSh, BSk | Hot or cold, large daily range | Very low rainfall (<250 mm desert, <500 mm steppe) | Sahara, Arabian Peninsula, central Australia |
| Temperate | Csa, Csb, Cfa, Cfb, Cwa | Mild winters, warm summers (10–25°C) | Moderate rainfall, seasonal in some areas | Europe, eastern US, China |
| Continental | Dfa, Dfb, Dfc, Dfd, Dwa, Dwb | Large seasonal range, cold winters (<0°C) | Moderate precipitation, often summer maximum | Canada, Russia, northern US |
| Polar | ET, EF | Warmest month <10°C | Low precipitation (often <250 mm/year) | Antarctica, Greenland |
| Highland | H (varies) | Varies with elevation | Varies with elevation | Andes, Himalayas |

*Table 12.8.2: Expanded classification table (selected subtypes):*

| Code | Description | Key characteristics |
|---|---|---|
| Af | Tropical rainforest | No dry season, heavy rainfall year-round |
| Am | Tropical monsoon | Short dry season, intense wet season |
| Aw | Tropical savanna | Distinct wet and dry seasons |
| BWh | Hot desert | Extremely dry, high temperatures |
| BWk | Cold desert | Dry with cooler winters |
| BSh | Hot semi-arid | Low rainfall, warm temperatures |
| BSk | Cold semi-arid | Low rainfall, cooler climate |
| Csa | Mediterranean (hot summer) | Dry summers, wet winters |
| Cfb | Marine west coast | Mild, wet year-round |
| Cfa | Humid subtropical | Hot summers, year-round precipitation |
| Dfa | Humid continental (hot summer) | Warm summers, cold winters |
| Dfc | Subarctic | Short cool summers, long cold winters |
| ET | Tundra | Short growing season |
| EF | Ice cap | Permanent ice, below freezing year-round |

*Table 12.8.3: Regional summary table*

| Region | Dominant climate types | Key characteristics |
|---|---|---|
| Equatorial regions | Af, Am | Warm, wet year-round |
| Subtropics | BWh, BSh, Csa | Dry or seasonal rain |
| Mid-latitudes | Cfa, Cfb, Dfa | Moderate temperatures |
| High latitudes | Dfc, ET, EF | Cold climates |
| Mountain regions | Highland | Elevation-dependent |

## Figure 12.2.2: Layered Paleoclimate Archives

Composite image showing five types of natural climate proxies that record past environmental conditions through layered growth patterns. From top to bottom, the image includes tree rings, varves (lake sediment layers), speleothems (cave mineral deposits), corals, and ice cores. Each section displays repeating bands or layers that represent changes over time, often on annual or seasonal scales.

Tree rings show alternating wide and narrow growth bands, typically representing yearly growth, where wider rings indicate favorable growing conditions and narrower rings indicate stress such as drought or cold. Varves show alternating sediment layers deposited in lakes, usually one pair of layers per year, with thicker or darker layers indicating higher sediment input. Speleothems display layered mineral deposits formed by dripping water in caves, with variations reflecting changes in precipitation and chemistry over time. Corals show banded growth structures, often annual, influenced by water temperature and chemistry. Ice cores show layered ice accumulation, where each layer represents a year or season, preserving information about past temperature, atmospheric composition, and precipitation.

These proxies collectively provide records spanning different time scales, from decades to hundreds of thousands of years.

## Figure 12.2.3: Reading Tree Rings

Diagram illustrating dendrochronology, the use of tree rings to reconstruct past events by matching patterns across different wood samples. Three horizontal wood sections are shown: a living tree, a dead tree, and a log from an older structure or ruins. Each section displays a sequence of growth rings, with alternating narrow and wide bands that represent yearly growth.

The living tree contains the most recent rings, including the outermost ring representing the current or most recent year. The dead tree shows an overlapping sequence of rings that match part of the living tree's pattern, allowing the timeline to be extended further into the past. The log from ruins contains an even older sequence of rings that overlaps with the dead tree's pattern, extending the record back further in time.

Distinctive ring patterns, such as unusually narrow or wide rings (shown as highlighted bands), align across the samples, demonstrating how overlapping sequences can be matched to create a continuous chronological record. This process allows scientists to build timelines that can extend hundreds to thousands of years into the past, with annual resolution (one ring per year).

## Figure 12.3.2: Global Temperature Reconstructions for past 2000 years.

Graph showing global temperature anomalies relative to the 1961–1990 average over the past 2000 years. The horizontal axis represents year from 0 to 2000, and the vertical axis shows temperature anomaly in degrees Celsius, ranging approximately from −0.7°C to +0.5°C. Three datasets are displayed: a proxy-based reconstruction (blue line) with an uncertainty range (shaded band), a set of discrete proxy estimates (green points), and modern instrumental data (red line).

From year 0 to about 1000, temperatures fluctuate slightly below the reference average, generally around −0.2°C to 0°C. Between about 1000 and 1800, temperatures trend cooler, reaching a minimum near −0.3°C to −0.4°C during the period often associated with the Little Ice Age. The uncertainty range widens further back in time, indicating less precise estimates.

After about 1900, the instrumental record (red line) shows a rapid warming trend, rising from around −0.2°C to approximately +0.4 to +0.5°C by 2000. This recent increase is steeper and exceeds the range of variability seen in the earlier part of the record. The proxy data (green points) generally align with the reconstructed trend but show variability around it.

*Table 12.8.4: Detailed data table (approximate temperature anomalies by period):*

| Time period (year) | Temperature anomaly (°C) | Characteristics |
|---|---|---|
| 0–500 | ~−0.2 to 0.0 | Slightly below average, moderate variability |
| 500–1000 | ~−0.2 to 0.0 | Relatively stable, small fluctuations |
| 1000–1400 | ~−0.2 to −0.3 | Gradual cooling trend |
| 1400–1800 | ~−0.3 to −0.4 | Coolest period (Little Ice Age) |
| 1800–1900 | ~−0.3 to −0.2 | Beginning of warming |
| 1900–2000 | ~−0.2 to +0.4 or +0.5 | Rapid modern warming |

*Table 12.8.5: Dataset comparison table:*

| Dataset | Time coverage | Data type | Key feature |
|---|---|---|---|
| Proxy reconstruction (blue) | 0–2000 | Smoothed reconstruction | Long-term trend with uncertainty band |
| Proxy data points (green) | 0–2000 | Individual estimates | Variability around reconstructed mean |
| Instrumental record (red) | ~1850–2000 | Direct measurements | Rapid recent warming trend |

*Table 12.8.6: Uncertainty characteristics table:*

| Time period | Uncertainty range (°C) | Interpretation |
|---|---|---|
| 0–500 | ±0.1 to ±0.2 | Moderate confidence |
| 500–1500 | ±0.2 to ±0.3 | Increasing uncertainty |
| 1500–1900 | ±0.3 to ±0.5 | Lower confidence in exact values |
| 1900–2000 | ±0.05 to ±0.1 | High confidence (instrumental data) |

## Figure 12.3.3: Global Surface Temperature Reconstruction during Holocene.

Graph showing global temperature anomalies relative to the 1961–1990 average over approximately the past 10,000 years (Holocene). The horizontal axis represents year from about −9000 (9000 BCE) to 2000 CE, and the vertical axis shows temperature anomaly in degrees Celsius, ranging roughly from −0.7°C to +0.5°C. A smoothed reconstruction (blue line) is shown with an uncertainty range (shaded band), and modern instrumental data (red line) appears near the present.

From about −9000 to −6000, temperatures rise from approximately −0.3°C to about +0.2 to +0.3°C, indicating early Holocene warming. Between about −6000 and −3000, temperatures remain relatively high and stable, generally around +0.2°C to +0.3°C. After about −3000, a gradual cooling trend begins, with temperatures declining toward 0°C and then below average.

From about −1000 to 1800 CE, temperatures fluctuate mostly between −0.1°C and −0.4°C, reaching a minimum near −0.4°C during the late pre-industrial period. Around 1900, the instrumental record shows a rapid warming, rising from about −0.2°C to approximately +0.4 to +0.5°C by 2000. The uncertainty band is wider in earlier periods, indicating less precise estimates, and narrows toward the present.

## Figure 12.3.5: Direct and Ice Core measurements of Carbon Dioxide.

Graph showing atmospheric carbon dioxide (CO2) concentration over the past 10,000 years, measured in parts per million (ppm). The horizontal axis represents time before 2005, from about 10,000 years ago to the present (0), and the vertical axis shows CO2 concentration ranging from about 250 ppm to over 380 ppm. Data points from multiple sources are plotted, with an inset zoom highlighting the most recent centuries.

From about 10,000 to 7,000 years ago, CO2 levels are relatively stable, fluctuating around 260–270 ppm. Between about 7,000 and 2,000 years ago, concentrations gradually increase to around 270–280 ppm. From about 2,000 years ago to the 18th century, CO2 remains fairly stable near 275–285 ppm with small variations.

Beginning around 1800, CO2 levels start to rise more rapidly, increasing from about 280 ppm to over 350 ppm by the late 20th century. By around 2005, concentrations reach approximately 380–390 ppm. The inset panel emphasizes this sharp modern increase, showing a steep upward trend compared to the relatively gradual changes over earlier millennia.

## Figure 12.3.8: Proxy-based Reconstruction of Past 800,000 years.

Four-panel graph showing linked climate changes over the past 800,000 years, with time on the horizontal axis from 0 (present) to 800 thousand years ago. The figure presents four aligned records: (a) oxygen isotope ratios (delta-18-O) from deep-sea benthic foraminifera, (b) atmospheric carbon dioxide concentration, (c) Antarctic air temperature anomaly, and (d) sediment reflectance in the Southern Ocean near Antarctica. Vertical shaded bands mark repeated glacial (colder) and interglacial (warmer) periods occurring roughly every 100,000 years.

In panel (a), delta-18-O values range from about 3.0 to 5.5 per mil. Lower values (about 3.0 to 4.0) indicate warmer periods with reduced global ice volume, while higher values (about 4.5 to 5.5) indicate colder periods with expanded ice sheets. The curve shows repeated sharp decreases (rapid warming into interglacials) followed by gradual increases (slow cooling into glacials), forming a sawtooth pattern across approximately eight major cycles.

In panel (b), atmospheric CO2 varies between about 180 ppm and 280 ppm. During glacial periods, CO2 is consistently low (about

180 to 200 ppm). During interglacial periods, CO2 rises to about 260 to 280 ppm. The timing of CO2 increases closely matches the warming transitions seen in the isotope and temperature records, with relatively rapid rises at the start of interglacials and slower declines during cooling phases.

In panel (c), Antarctic temperature anomaly ranges from approximately –10°C during cold glacial maxima to near 0°C during warm interglacials. Temperature increases rapidly during transitions into interglacial periods (warming of roughly 8 to 10°C), then gradually decreases over tens of thousands of years. The temperature curve closely follows the CO2 record and inversely tracks the isotope record (higher temperatures correspond to lower delta-18-O values).

In panel (d), sediment reflectance ranges from about 10% to 40% and serves as a proxy for ocean conditions such as productivity and stratification. Higher reflectance values (around 25% to 40%) generally occur during warmer interglacial periods, while lower values (around 10% to 20%) occur during colder glacial periods. The pattern shows variability within each cycle but broadly aligns with temperature and CO2 changes.

Across all panels, the records demonstrate strong coupling: warm interglacial periods are characterized by lower delta-18-O values (~3–4), higher CO2 (~260–280 ppm), higher Antarctic temperatures (near 0°C anomaly), and higher sediment reflectance (~25–40%). Cold glacial periods show the opposite pattern: higher delta-18-O (~4.5–5.5), lower CO2 (~180–200 ppm), lower temperatures (~–8 to –10°C), and lower reflectance (~10–20%). The transitions into warm periods are relatively abrupt (on the order of thousands of years), while cooling into glacial conditions is more gradual (tens of thousands of years).

The figure illustrates about eight major glacial-interglacial cycles over 800,000 years, each lasting roughly 100,000 years, with consistent amplitude and timing relationships among ice volume, greenhouse gases, temperature, and ocean conditions.

## Figure 12.4.1: Tectonic Plates

Global map showing the major tectonic plates of Earth and their boundaries. The continents and oceans are divided into large plates that move relative to one another. The largest plate is the Pacific Plate, which covers much of the Pacific Ocean basin. Other major plates include the North American, South American, African, Eurasian, Antarctic, Australian, and Indian plates. Several smaller plates are also shown, including the Nazca, Cocos, Caribbean, Philippine, Arabian, Somali, Scotia, and Juan de Fuca plates.

Plate boundaries are marked with different line types indicating interactions. Divergent boundaries (where plates move apart) are primarily located along mid-ocean ridges, such as the Mid-Atlantic Ridge between the North American and Eurasian plates in the north and between the South American and African plates in the south. Convergent boundaries (where plates move toward each other) occur along subduction zones, such as where the Nazca Plate meets the South American Plate along the west coast of South America, and where the Pacific Plate meets surrounding plates along the Pacific “Ring of Fire.” Transform boundaries (where plates slide past each other) occur in locations such as the boundary between the Pacific Plate and North American Plate along western North America.

The map shows that plate boundaries often align with major geological features: mountain ranges (for example, the Himalayas at the boundary of the Indian and Eurasian plates), ocean trenches (for example, along the western Pacific), and mid-ocean ridges. Plate sizes vary widely, with major plates spanning thousands of kilometers, while smaller plates occupy more localized regions.

## Figure 12.4.2: Solar Cycle

Graph showing variations in solar activity and solar irradiance from about 1975 to 2005. The horizontal axis represents year, and the vertical axis shows solar irradiance in watts per square meter ($W/m^2$), ranging approximately from 1365 to 1367 $W/m^2$. Multiple lines represent different indicators of solar activity, including total solar irradiance (daily and annual averages), sunspot observations, solar flare index, and 10.7 cm radio flux.

The data show repeating cycles approximately every 10 to 11 years. Solar irradiance varies slightly, with peaks around 1366.5 to 1367 $W/m^2$ during periods of high solar activity and lows near 1365.5 $W/m^2$ during quieter periods. These peaks occur around the early 1980s, around 1990, and around 2000.

Sunspot observations, solar flare index, and radio flux all rise and fall in sync with irradiance, indicating that higher solar activity corresponds to slightly increased solar energy output. The daily irradiance (shown with large variability) fluctuates rapidly, while the smoothed annual values show a clearer cyclic pattern. Overall, the amplitude of irradiance variation is small (about 1 to 1.5 W/

$m^2$, or less than 0.1% of total output), despite larger variations in activity indicators.

*Table 12.8.7: Detailed data table (solar cycle characteristics):*

| Parameter | Approximate range | Behavior |
|---|---|---|
| Solar irradiance ($W/m^2$) | ~1365.5–1367 | Small cyclic variation |
| Cycle length | ~10–11 years | Repeating solar cycle |
| Irradiance variation | ~1–1.5 $W/m^2$ | Less than 0.1% change |
| Peak years (approx.) | ~1980, ~1990, ~2000 | High solar activity |
| Minimum years (approx.) | ~1976, ~1986, ~1996, ~2005 | Low solar activity |

## Figure 12.4.3: Milankovitch Cycles

Diagram illustrating the three main components of Milankovitch cycles, which describe long-term variations in Earth's orbit and orientation that influence climate. The image shows the Sun at the center and Earth orbiting along an elliptical path. Three labeled processes are indicated: eccentricity (E), tilt or obliquity (T), and precession (P).

Eccentricity (E) refers to changes in the shape of Earth's orbit from more circular to more elliptical. The diagram shows an elongated orbit, indicating that the distance between Earth and the Sun varies over time. This cycle occurs on timescales of about 100,000 years (with additional longer cycles around 400,000 years) and affects the overall distribution of solar energy received by Earth.

Tilt (T), also called obliquity, refers to changes in the angle of Earth's axis relative to its orbital plane. The diagram shows Earth's axis tilted relative to vertical, with arrows indicating variation in this angle. The tilt varies between about 22.1° and 24.5° over a cycle of approximately 41,000 years. Greater tilt increases seasonal contrast (warmer summers and colder winters), especially at higher latitudes.

Precession (P) refers to the wobble of Earth's rotational axis, changing the orientation of the axis over time. The diagram shows circular arrows around the axis, indicating this slow rotation. Precession operates on timescales of about 19,000 to 23,000 years and affects the timing of seasons relative to Earth's position in its orbit (for example, whether summer occurs when Earth is closer to or farther from the Sun).

Together, these three cycles alter the distribution and intensity of solar radiation received at different latitudes and seasons, contributing to long-term climate patterns such as glacial and interglacial cycles.

## Figure 12.4.4: Milankovitch Cycles and Past Glaciation

Multi-line graph showing variations in Earth's orbital parameters and their relationship to climate over the past 1,000,000 years (1 million years). The horizontal axis represents time from 1,000 thousand years ago (kyr ago) to the present ("Now"). Five stacked panels display: precession, obliquity, eccentricity, solar forcing at 65°N during summer, and stages of glaciation (temperature state). Vertical shaded bands mark alternating colder (glacial) and warmer (interglacial) periods.

The top panel shows precession, with rapid oscillations occurring approximately every 19, 22, and 24 thousand years. The signal is highly variable, with frequent peaks and troughs indicating changes in the timing of seasons relative to Earth's orbit.

The second panel shows obliquity (axial tilt), varying smoothly with a cycle of about 41,000 years. The amplitude remains relatively consistent, reflecting changes in Earth's tilt between about 22° and 24.5°, which influences seasonal contrast.

The third panel shows eccentricity, with longer cycles of about 95, 125, and 400 thousand years. The variations are slower and lower frequency, indicating changes in the shape of Earth's orbit from more circular to more elliptical.

The fourth panel shows solar forcing at 65°N during summer, which combines the effects of precession, obliquity, and eccentricity. This curve fluctuates on multiple timescales, with both high-frequency variations (tens of thousands of years) and lower-frequency modulation (hundreds of thousands of years). Peaks represent stronger summer solar input at high northern latitudes, while

troughs represent weaker input.

The bottom panel shows stages of glaciation, with a temperature-related curve labeled from “cold” to “hot.” This curve shows repeated glacial-interglacial cycles, with cold periods lasting longer and warm periods appearing as shorter peaks. The timing of these climate changes broadly aligns with variations in solar forcing, especially the longer eccentricity-modulated cycles.

Overall, the figure illustrates how orbital variations influence solar energy distribution and are linked to the timing of ice ages, with dominant glacial cycles occurring roughly every 100,000 years.

*Table 12.8.8: Detailed data table (orbital cycles):*

| Parameter | Cycle lengths (kyr) | Frequency pattern | Primary effect |
|---|---|---|---|
| Precession | ~19, 22, 24 | High frequency | Seasonal timing changes |
| Obliquity | ~41 | Moderate frequency | Controls seasonal intensity |
| Eccentricity | ~95, 125, 400 | Low frequency | Modulates orbital shape |

*Table 12.8.9: Solar forcing table:*

| Feature | Timescale influence | Description |
|---|---|---|
| High-frequency variation | ~20–40 kyr | Driven by precession and obliquity |
| Low-frequency modulation | ~100–400 kyr | Controlled by eccentricity |
| Peak forcing | — | Strong summer insolation at 65°N |
| Minimum forcing | — | Weak summer insolation |

*Table 12.8.10: Glaciation characteristics table:*

| Feature | Approximate value | Description |
|---|---|---|
| Cycle duration | ~100 kyr | Dominant glacial cycle |
| Cold period length | ~80–90 kyr | Long glacial intervals |
| Warm period length | ~10–20 kyr | Short interglacial peaks |
| Temperature pattern | Asymmetric | Gradual cooling, rapid warming |

## Figure 12.4.5c: Volcanic Eruptions and Solar Transmission

Graph showing the percentage of solar radiation transmitted through Earth’s atmosphere from about 1958 to 2008. The horizontal axis represents year, and the vertical axis shows solar radiation transmitted in percent, ranging from about 70% to 100%. A relatively steady baseline near 92–94% is visible, with several sharp decreases corresponding to major volcanic eruptions.

Most of the record shows small fluctuations around about 93%, indicating relatively stable atmospheric transmission. However, three major dips stand out. Around 1982, transmission drops sharply to about 78–80%, corresponding to the eruption of El Chichón (17.4°N, Volcanic Explosivity Index (VEI) 5). Around 1991, an even larger drop occurs to about 75–80%, associated with the eruption of Mount Pinatubo (15.1°N, VEI 6). Smaller dips are also visible around 1963 (Agung, 8.3°S, VEI 4) and around 1974–1975 (Fuego, 14.5°N, VEI 4), with transmission decreasing to roughly 90–92%.

After each eruption, solar transmission gradually recovers over several years as volcanic aerosols settle out of the atmosphere. The largest eruptions produce the deepest and longest-lasting reductions in transmitted solar radiation.

# Figure 12.5.5: Trends in Sea Ice

Three-panel figure showing sea ice extent trends from 1979 to about 2020–2021 for the Northern Hemisphere (top), global total (middle), and Southern Hemisphere (bottom). The horizontal axes show year, and the vertical axes show sea ice extent in million square kilometers. Each panel includes annual bars, a long-term average line (1981–2010), and a linear trend line indicating change over time.

In the Northern Hemisphere panel (September, annual minimum), sea ice extent declines from about 7.5–7.8 million km² in the late 1970s to about 4.0–4.5 million km² by 2020. The 1981–2010 average is about 6.41 million km². The trend shows a decrease of about −12.79% per decade, equivalent to approximately −0.82 million km² per decade. Year-to-year variability is present, with some higher years in the 1980s–1990s and notably low values after 2007.

In the Global sea ice panel (monthly data, 1979–2021), total sea ice extent fluctuates seasonally between roughly 16 and 27 million km². The long-term average is about 23.25 million km². The overall trend is slightly negative, about −2.1% per decade (approximately −0.51 million km² per decade). Seasonal cycles dominate the variability, but a gradual decline is visible over time.

In the Southern Hemisphere panel (March, annual minimum), sea ice extent varies between about 3.0 and 5.2 million km². The 1981–2010 average is about 4.03 million km². The trend shows a small increase of about +1.74% per decade (approximately +0.07 million km² per decade), with noticeable variability and some higher values in the 2000s–2010s, followed by lower values in recent years.

Overall, the figure shows a strong decline in Arctic (Northern Hemisphere) sea ice, a slight increase in Antarctic (Southern Hemisphere) sea ice over much of the period, and a small net decline in global sea ice, with variability driven by seasonal cycles and regional differences.

# Figure 12.5.7: Mountain Glacier Decline

Graph showing cumulative mass change of reference glaciers across multiple global regions from about 1950 to 2020. The horizontal axis represents year, and the vertical axis shows cumulative mass change in meters of water equivalent (m w.e.), ranging approximately from +5 to −35. Values are relative to a baseline around 1976, where most curves cross near zero. Multiple colored lines represent different regions, and a dashed black line shows the global mean across regions.

From 1950 to the mid-1970s, most regions show small fluctuations around zero, generally within about +2 to −2 m w.e., indicating relatively stable glacier mass. After about 1980, nearly all regions show a consistent downward trend, indicating cumulative mass loss.

By around 2000, most regions have lost between about −5 and −15 m w.e. After 2000, the rate of loss accelerates in many regions. By 2020, cumulative losses vary widely by region:

- Western Canada and USA show the largest loss, reaching approximately −30 to −35 m w.e.
- Central Europe also shows large losses, around −25 to −30 m w.e.
- Alaska, Asia Central, and Arctic Canada North show losses of roughly −15 to −25 m w.e.
- Scandinavia, Southern Andes, and Svalbard and Jan Mayen show somewhat smaller but still substantial losses, typically around −10 to −20 m w.e.
- Asia North and the Caucasus and Middle East show losses closer to −10 to −15 m w.e.

The dashed global mean declines from near 0 in 1976 to about −20 m w.e. by 2020, indicating widespread and consistent glacier mass loss across regions. The curves show increasing divergence after 1990, reflecting regional differences in climate and glacier response, but all trends are negative after the late 20th century.

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*Table 12.8.11: Detailed data table (approximate cumulative mass loss by region, ~2020):*

| Region | Cumulative mass change (m w.e.) | Relative magnitude |
|---|---|---|
| Western Canada and USA | ~−30 to −35 | Very high loss |
| Central Europe | ~−25 to −30 | Very high loss |

| Region | Cumulative mass change (m w.e.) | Relative magnitude |
|---|---|---|
| Alaska | ~−20 to −25 | High loss |
| Asia Central | ~−18 to −25 | High loss |
| Arctic Canada North | ~−15 to −22 | High loss |
| Svalbard and Jan Mayen | ~−15 to −20 | Moderate-high loss |
| Southern Andes | ~−12 to −20 | Moderate-high loss |
| Scandinavia | ~−10 to −15 | Moderate loss |
| Caucasus and Middle East | ~−10 to −15 | Moderate loss |
| Asia North | ~−10 to −12 | Lower (but still significant) loss |

## Figure 12.5.8: Greenland and Antarctica Ice Sheet Loss

Line graph showing cumulative ice mass change for Antarctica and Greenland from about 1992 to 2024. The horizontal axis represents year, and the vertical axis shows cumulative mass change in billion metric tons, ranging from about +500 to −6000. A dashed horizontal line at 0 marks no net change. Two main datasets are shown for each region: a combined estimate (bold lines) and a reference dataset (lighter lines), with shaded bands indicating uncertainty.

At the start of the record (early 1990s), both Antarctica and Greenland are near 0 change, indicating little net mass loss relative to the baseline. Over time, both regions show increasingly negative values, indicating cumulative ice loss.

Greenland (blue lines) shows a steep and accelerating decline. By about 2005, Greenland has lost roughly −500 to −1000 billion tons. Loss accelerates after 2010, reaching about −3000 billion tons by around 2015, and approximately −4500 to −5500 billion tons by 2024. The trend is strongly downward, with short-term variability but a clear long-term acceleration.

Antarctica (red lines) also shows a net loss, but at a slower rate. By about 2005, losses are around −200 to −400 billion tons. By 2015, cumulative loss reaches roughly −1000 to −1500 billion tons. By 2024, Antarctica has lost approximately −2000 to −2500 billion tons. The trend is downward but less steep than Greenland's.

The uncertainty bands widen slightly over time, but both datasets consistently show negative trends. Greenland contributes the larger share of total ice mass loss over this period, with losses roughly two to three times greater than Antarctica by the end of the record.

## Figure 12.5.9: Oceanic Heat Content

Graph showing global ocean heat content (0–2000 meters depth) from about 1960 to 2025. The horizontal axis represents year, and the vertical axis shows heat content in units of 10^22 joules, ranging from about −10 to +35. Multiple lines represent different averaging methods: a long-term smoothed trend (blue), yearly averages (black), and a recent 3-month average (red).

From about 1960 to the early 1970s, ocean heat content is relatively low, fluctuating between about −10 and −5 × 10^22 joules. Between the 1970s and 1990, values gradually rise to around 0 to +5 × 10^22 joules, with some short-term variability.

After about 1990, the increase becomes more pronounced. By around 2005, heat content reaches approximately +10 × 10^22 joules. The rate of increase accelerates further after 2010, with values rising to about +20 × 10^22 joules by the mid-2010s and continuing upward.

By 2024–2025, ocean heat content reaches approximately +30 to +33 × 10^22 joules, representing the highest values in the record. The red line (recent 3-month average) shows short-term variability but follows the same upward trend. The blue line (smoothed trend) highlights a steady and accelerating increase over time.

Overall, the graph shows a long-term increase in ocean heat content, with the majority of warming occurring after 1990 and especially after 2010.

## Figure 12.5.10: Global Mean Sea Level Change

Graph showing global mean sea level change from about 1880 to 2020, with a secondary pie chart illustrating contributions to sea level rise from 2005 to 2015. The horizontal axis represents year, and the vertical axis shows global mean sea level in millimeters (mm), ranging from about −50 mm to +250 mm relative to a baseline in the late 19th century.

The main time series shows a steady increase in sea level over time. From about 1880 to 1920, sea level rises gradually from approximately −20 mm to about +40 mm. Between 1920 and 1970, the increase continues more steadily, reaching around +100 mm. From 1970 onward, the rate of rise accelerates. By about 2000, sea level reaches approximately +150 mm, and by 2020 it approaches about +240 to +250 mm.

Two data sources are shown: tide gauge measurements (blue line with shaded uncertainty band) covering the full record, and satellite altimetry (red line) from the early 1990s onward. The satellite data show a steeper rise compared to earlier decades, confirming an acceleration in sea level increase.

The inset pie chart shows contributions to sea level rise during 2005–2015. Thermal expansion of seawater accounts for about 40% of the total rise. Melting glaciers contribute about 17%, Greenland ice sheet loss contributes about 21%, Antarctic ice sheet loss contributes about 12%, and groundwater depletion contributes about 11%.

Overall, the figure shows a long-term rise in global sea level with clear acceleration in recent decades, driven by both ocean warming and increasing ice loss.

## Figure 12.5.11: The Keeling Curve

Line graph showing atmospheric carbon dioxide (CO2) concentrations measured at Mauna Loa Observatory from about 1960 to 2020. The horizontal axis represents year, and the vertical axis shows CO2 concentration in parts per million (ppm), ranging from about 315 ppm to over 410 ppm.

The graph shows a clear long-term upward trend in CO2 levels. In 1960, concentrations are approximately 315–317 ppm. By 1980, levels rise to about 335–340 ppm. Around 2000, CO2 reaches roughly 370 ppm. By 2010, concentrations increase further to about 390 ppm, and by 2020 they exceed 410 ppm.

Superimposed on the long-term increase is a regular seasonal cycle, visible as a repeating sawtooth pattern each year. The seasonal variation is typically about 5–8 ppm, with peaks occurring in late spring and lows in early autumn, reflecting natural cycles of plant growth and decay in the Northern Hemisphere.

The overall trend shows an accelerating increase, with the rate of rise growing from about 1 ppm per year in the 1960s to about 2–3 ppm per year in recent decades. The data are attributed to long-term monitoring programs by scientific institutions, and the record is commonly known as the Keeling Curve.

## Figure 12.6.4: U.S. Electricity Generation Sources

Chart showing sources of U.S. electricity generation in 2023, totaling about 4.18 trillion kilowatt-hours. The graphic divides electricity generation into major categories with percentage shares. Natural gas is the largest source at about 43.1% of total generation. Renewable energy accounts for about 21.4%, nuclear power contributes about 18.6%, coal provides about 16.2%, and petroleum and other sources account for about 0.8%.

Within the renewable category (21.4%), the breakdown includes wind at about 10.2%, hydropower at about 5.7%, solar at about 3.9%, biomass at about 1.1%, and geothermal at about 0.4%.

Overall, fossil fuels (natural gas, coal, and petroleum) together account for about 60.1% of electricity generation, while non-fossil sources (renewables plus nuclear) make up about 40.0%.

## Figure 12.6.5b: Hydroelectric Dams

Diagram of a hydroelectric power plant showing how water stored in a reservoir is used to generate electricity. Water is held

behind a dam in a reservoir at a higher elevation. It enters the system through an intake structure and flows into a large pipe called a penstock. The water travels downward through the penstock, gaining speed due to gravity.

At the lower end of the system, the flowing water turns a turbine. The turbine is connected to a generator located inside a powerhouse. As the turbine spins, it drives the generator to produce electricity. Electrical power is then transmitted away from the plant through power lines. After passing through the turbine, the water exits the system and flows into a downstream river.

The diagram shows the sequence: reservoir → intake → penstock → turbine → generator (in powerhouse) → transmission lines → river. The system converts gravitational potential energy of stored water into mechanical energy (turbine rotation) and then into electrical energy.

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