1 BASICS

WHAT IS CLIMATE?

Robert Heinlein wrote that "[c]limate is what we expect, weather is what we get."[1] What we expect is *typical* weather, but the weather at a given place, on a given day, can be very atypical.

A dictionary definition of climate is "the average course or condition of the weather at a place usually over a period of years as exhibited by temperature, wind velocity, and precipitation" (Merriam-Webster online, http://www.merriam-webster.com/dictionary/climate). Here "average" refers to a time average. In many reference works, climatological averages are defined to be taken over 30 years, a definition that probably has to do more with the human life span than with any physical time scale.

Simple 30-year averages are not enough because they hide important variability, such as differences between summer and winter. Because of the great importance of the seasonal cycle, climatological averages are often specified for particular months of the year; we might discuss the climatological average precipitation rate for New York City, averaged over 30 Julys, or 30 Januarys.

Climate varies geographically, most obviously between the tropics and the poles. There are also important climate variations with longitude, at a given latitude.[2] For example, the Sahara Desert and the jungles of southeast Asia have very different climates, even though they are at the same latitude. Climate also varies strongly with surface elevation; near the surface, the temperature and water vapor concentration of the air generally decrease upward, and the wind speed often increases upward.

The climate of the Earth as a whole changes with time, for example when ice ages come and go. In fact, the climate of the Earth as a whole is changing right now due to rapid, anthropogenically produced changes in the composition of the atmosphere.

The climate parameters of greatest importance to people are precipitation and temperature. The next few items on the list would include wind speed and direction, and cloudiness. Scientists are interested in a much longer list of parameters, of course.

The "average course or condition of the weather," mentioned in the dictionary definition of climate, includes not only simple averages, such as the climatological January mean surface air temperature, but also, importantly, *statistics that characterize the fluctuations and variations of the climate system*. Variability is of great and even primary interest. Predictions are all about change. Examples of important variations that are aspects of climate include the following:

The seasonal variations of surface air temperature (and many other things)

Systematic day-night temperature differences

- The tendency of thunderstorms to occur in late afternoon in many places
- The frequency of snow storms
- The occurrence, every few years, of "El Niño" conditions, which include unusually warm sea surface temperatures in the eastern tropical Pacific Ocean

The list could easily be extended. Variations like these are important aspects of the climate state. All of them can be described by suitably concocted statistics; for example, we can discuss the average daily minimum and maximum near-surface air temperatures.

It is important to distinguish between "forced" and "free" variations. Forced variations include the day-night and seasonal changes mentioned above, which are externally driven by local changes of solar radiation that are associated with the Earth's rotation on its axis and its orbital motion around the Sun, respectively. Volcanic eruptions can also force climate fluctuations that sometimes last for years. On the other hand, storms and El Niños are examples of unforced or "free" variations that arise naturally through the internal dynamics of the climate system.

For the reasons outlined above, I would modify the dictionary definition to something like this: "Climate is the (in principle, infinite) collection of statistics based on the evolving, geographically distributed state of the atmosphere, including not only simple averages but also measures of variability on a range of time scales from hours to decades."

Since "weather" refers to the state of the atmosphere, the definitions given above make climate appear to be a property of the atmosphere alone. Climate scientists don't think of it that way, though, because any attempt to understand what actually determines the state of the climate, and what causes the climate to change over time, has to take into account the crucial roles played by the ocean (including marine biology), the land surface (including terrestrial biology), and the continental ice sheets. These, together with the atmosphere, make up the four primary components of what is often called the "climate system."

The ocean is about 400 times more massive than the atmosphere and has a heat capacity more than a thousand times larger.[3] When the ocean says "Jump," the atmosphere asks "How high?" Nevertheless, the thin, gaseous atmosphere exerts a powerful influence on the climate. How can the relatively puny atmosphere play such a major role in the much larger climate system?

The explanation has two parts. First, the atmosphere serves as an outer skin, standing between the other components of the climate system and space. As a result, the atmosphere is in a position (so to speak) to regulate the all-important exchanges of energy between the Earth and space, which take the forms of solar radiation coming in and infrared radiation going out.

The second reason is that the atmosphere can transport energy, momentum, and other things from place to place much faster than any other component of the climate system. Typical wind speeds are hundreds or even thousands of times faster than the speeds of ocean currents, which are in turn much faster than the ponderous motions of the continents. The atmosphere (and ocean) can also transport energy through the pressure forces exerted by rapidly propagating fluid-dynamical waves of various kinds.[4]

The climate system is of course governed by the laws of physics. Its behavior can be measured in terms of its physical properties, analyzed in terms of its physical processes, and predicted using physical models. It is influenced by a variety of "external" parameters that are (almost) unaffected by processes at work inside the climate system. These external parameters include the size, composition, and rotation rate of the Earth; the geographical arrangement of oceans, continents, mountain ranges, and so on; the geometry of the Earth's orbit around the Sun; and the amount and spectral distribution of the electromagnetic radiation emitted by the Sun.[5]

Over the past few decades, the possibility of ongoing and future anthropogenic climate change has been widely recognized as a major scientific and societal issue, with huge economic ramifications. As a result, the physical state of the climate system is now being intensely monitored, like the health of a patient with worrisome symptoms. Ongoing changes are being diagnosed. The future evolution of the system is being predicted, using rapidly improving physically based models that run on the fastest computers in the world.

THE COMPOSITION OF THE ATMOSPHERE

The atmosphere is big. Its total mass is about 5×10^{21} g.

The most abundant atmospheric constituents are nitrogen and oxygen. They are very well mixed throughout almost the entire atmosphere, so that their relative concentrations are essentially constant in space and time.

Ozone and water vapor are "minor" but very important atmospheric constituents that are *not* well mixed, because they have strong sources and sinks inside the atmosphere. Ozone makes up less than one millionth of the atmosphere's mass, but that is enough to protect the Earth's life from deadly solar ultraviolet (UV) radiation. Water vapor is only about a quarter of 1% of the atmosphere's mass, but its importance for the Earth's climate, and for the biosphere, would be hard to exaggerate, and it will be discussed at length throughout this book.

The term "dry air" refers to the mixture of atmospheric gases other than water vapor. As shown in Table 1.1, dry air is a mixture of gases. Each gas approximately obeys the ideal gas law, which can be written for a particular gas, denoted by subscript i, as

$$p_i V = N_i k T \tag{1.1}$$

Here p_i is the partial pressure of the gas, V is the volume under consideration, N_i is the number of particles, k is Boltzman's constant, and T is the temperature. If the gas is in thermal equilibrium, then the temperature will be the same for all gases in the mixture. This is an excellent assumption for dry air, valid up to at least 100 km above the surface. We can write $N_i k = n_i R^*$, where n_i is the number of moles and R^* is the universal gas constant.

Table 1.1
A partial list of the well-mixed gases that make up "dry air," in order of their mass fraction of the atmosphere. Oxygen is present in significant quantities only because of the existence of life on Earth. Three of the six leading constituents are noble gases (argon, neon, and helium). Additional gases (not listed) are present in smaller amounts.

Gas	*Molecular form*	*Molecular mass g mol^{-1}*	*Volume fraction ppmv*	*Mass fraction of the "dry" portion of the atmosphere*
Nitrogen	N_2	28.0	781,000	0.755
Oxygen	O_2	32.0	209,000	0.231
Argon	Ar	39.4	9,340	0.0127
Carbon dioxide	CO_2	44.0	390	5.92×10^{-4}
Neon	Ne	20.2	18	1.26×10^{-5}
Helium	He	4.0	5	6.90×10^{-7}
Methane	CH_4	16.0	2	1.10×10^{-7}

The total mass of the gas, M_i, satisfies $n_i = \frac{M_i}{m_i}$, where m_i is the molecular mass. Substituting, we find that

$$p_i = \rho_i \frac{R^*}{m_i} T, \tag{1.2}$$

where $\rho_i \equiv \frac{M_i}{M_i}$ is the density.

The temperature is a measure of the kinetic energy of the random molecular motions. It normally decreases with height in the lower atmosphere, although, as discussed later, it actually increases upward at greater heights. The range of temperatures encountered throughout most of the atmosphere is roughly 200 K to 300 K.[6]

The pressure of a gas is, by definition, the normal component of the force (per unit area) exerted by the moving molecules. In an ideal fluid, the pressure at a point is the same in all directions.

You have probably experienced the increased pressure that the water exerts on your ears (and the rest of your body) at the bottom of a swimming pool. What you are feeling is the weight (per unit horizontal area) of the water above you. At greater depths, there is more water above, it pushes down on you more heavily, and the water pressure increases as a direct result. The density of the water, ρ_{water}, is very nearly constant, so the pressure at a given depth, D, is given by $p = \rho_{\text{water}} gD$, where g is the acceleration of gravity, which is about 9.8 m s^{-2} near the Earth's surface.[7]

Similarly, the air pressure at a given height is very nearly equal to the weight (per unit horizontal area) of the air above. In a "high-pressure" weather system, you are buried under a thicker (more massive) layer of air. With a low-pressure system, the layer of air is thinner. This "hydrostatic" relationship applies to each gas separately because the weights of the gases simply add. For a particular gas, denoted by subscript i, the hydrostatic relationship can be expressed in differential form by

$$\frac{\partial p_i}{\partial z} = -\rho_i g, \tag{1.3}$$

where z is height. The minus sign appears in Equation (1.3) because the pressure increases downward while height increases upward. We use a *partial* derivative of p with respect to z in (1.3) because the pressure also depends on horizontal position and time. The hydrostatic equation, (1.3), expresses a balance between two forces, namely the downward weight of the air and the upward pressure force that arises from the upward decrease of pressure. The balance is not exact. It is an approximation, which is another way of saying that it has an error because something has been neglected. That something is the actual acceleration of the air in the vertical direction. Further discussion is given in Chapter 3.

By combining the ideal gas law with the hydrostatic equation, we find that

$$\frac{1}{p_i}\frac{\partial p_i}{\partial z} = -\frac{m_i g}{R^* T}. \tag{1.4}$$

As you probably know, g decreases upward in proportion to the square of the distance from the center of the Earth. The Earth's atmosphere is very thin, though, so the variations of g with height inside the atmosphere are negligible for most purposes, and we will neglect them here. Suppose that the temperature varies slowly with height. Treating it as a constant,[8] we can integrate both sides of (1.4) to obtain

$$p_i(z) = \left(p_i\right)_S e^{-\frac{z}{H_i}}, \tag{1.5}$$

where the surface height is taken to be zero, $\left(p_i\right)_S$ is the surface partial pressure of gas i, and

$$H_i \equiv \frac{R^* T}{m_i g} \tag{1.6}$$

is called the "scale height" of the gas. Equation (1.5) says that the partial pressure decreases upward exponentially away from the surface, at a rate that depends on the molecular mass of the gas and also on the temperature, which, again, is the same for all of the gases in the mixture. By putting in some numbers, you will find that a typical value of the scale height of nitrogen in the Earth's atmosphere is about 8,000 m. According to Equation (1.5), the partial pressure of nitrogen decreases upward by a factor of e in 8,000 m.

How heavy is a column of air? The total pressure (due to the weight of all atmospheric constituents) is typically about 100,000 Pa near sea level, where a Pa (pascal) is defined to be a newton (Nt) per square meter. For comparison, the weight of a typical car is about 20,000 Nt, so the weight of an air column per square meter is roughly equivalent to the weight of five cars piled on top of each other, over one square meter of a junk yard. That's pretty heavy. The total pressure at an altitude of 12 km is about 20,000 Pa, roughly 5 times less than the surface pressure.

By pushing the ideas presented above just a bit further, you should be able to show that the partial density of a gas also decreases upward exponentially, following a formula very similar to (1.4). The total density of the air near sea level is typically about 1.2 kg m^{-3}. The total density at an altitude of 12 km is about 5 times less. As an example, the density of atmospheric oxygen decreases exponentially upward, which is why hiking is more challenging at higher altitudes.

As shown in Table 1.1, the various atmospheric gases have different molecular masses, and as a result they have different scale heights. This suggests that the relative concentrations of the various gases should vary with height, so that the heaviest species would be more concentrated near the ground and the lighter species more concentrated higher in the atmospheric column. The process that could lead to such a result is called "diffusive separation" because it depends on the microscopic shaking due to molecular motions, in the presence of gravity. Diffusive separation is a real process, but it is negligible in the lower atmosphere because the turbulent winds act like a powerful mixer that homogenizes the blend of gases. The result is that, except for ozone, the concentrations (*not* the densities) of the gases that make up dry air, for example, nitrogen, oxygen, argon, and carbon dioxide, are observed to be nearly homogeneous, both horizontally and vertically, up to an altitude of about 100 km. Above that level, diffusive separation does become noticeable.

As an example, Figure 1.1 shows the variations of density, pressure, and temperature with height, from the surface to an altitude of 50 km, based on the U.S. Standard Atmosphere (http://modelweb.gsfc.nasa.gov/atmos/us_standard.html).[9]

Vertical profiles are called "soundings," a term borrowed from oceanography. The approximately exponential upward decreases of density and pressure are obvious. The temperature decreases upward for the first 10 km or so, remains almost constant for the next 10 km,

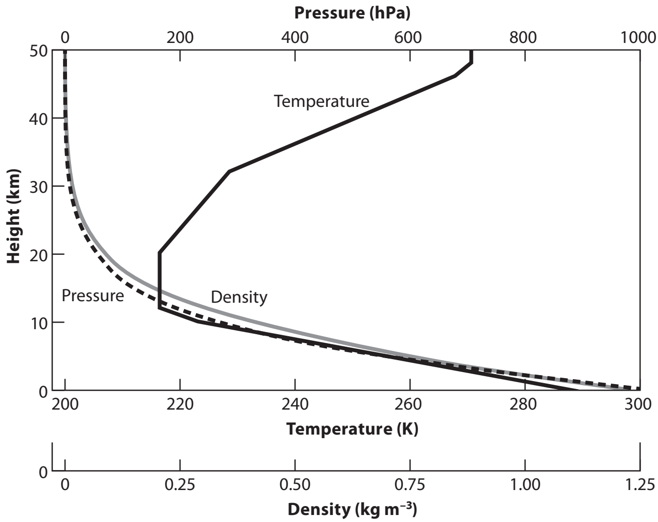

Figure 1.1. Typical variations of density, pressure, and temperature with height, from the surface to an altitude of 50 km, which is in the upper stratosphere.

The exponential upward decreases of density and pressure are clearly evident. The temperature variations are quite different; near the surface it decreases upward, but above about the 20 km level it actually increases upward, for reasons discussed in the text. The scales for density and pressure start at zero, while the scale for temperature starts at 200 K. This figure is based on the U.S. Standard Atmosphere.

and then begins to increase upward. This vertical distribution of temperature will be explained later.

Water vapor is an important exception to the rule that the atmosphere is well mixed. There are two main reasons for this. First, water vapor enters the atmosphere mostly by evaporation from the ocean, so it has a tendency to be concentrated near the surface. More importantly, water vapor can condense to form liquid or ice, which then falls out of the atmosphere as rain or snow. Condensation happens when the actual concentration of vapor exceeds a saturation value, which depends on temperature. In effect, the water vapor concentration is limited to be less than or equal to the saturation value. At the colder temperatures above the surface, saturation occurs more easily, and the water concentration decreases accordingly. The fact that water can change its phase in our atmosphere is critically important for the Earth's climate. Much additional discussion is given in later chapters.

In a warm place near the surface, water vapor can be as much as 2% of the air by mass. You would drip with sweat in a place like that. In a cold place, the air might contain 0.1% water vapor by mass; your skin would tend to dry out and crack under those conditions. In the stratosphere, the mass fraction of water vapor is typically just a few parts per million.[10]

In addition to the various gases discussed above, each cubic centimeter of air contains many, sometimes hundreds, of tiny liquid and solid particles called aerosols. Aerosols include dust, much of which comes from deserts, where it is scoured away from the land surface by the drag force associated with the winds. The dust can be carried over vast distances. For example, Asian dust is carried eastward across the Pacific to California, and Saharan dust is carried westward across the Atlantic to Florida and eastern South America. Another important type of aerosol is sea salt. The spray whipped up by winds near the sea surface evaporates, leaving behind tiny salt particles, suspended in the air, that can then be carried long distances by the winds. Pollutants emitted by human activities are another major source of aerosols. These anthropogenic aerosols include sulfur compounds and also carbon compounds such as soot.

The most obvious liquid and solid particles in the atmosphere are the liquid water drops and ice crystals that make up clouds. As a matter of terminology, however, atmospheric scientists conventionally reserve the term "aerosol" to refer to the wide variety of *noncloud* liquid and solid particles in the air. Cloud particles range in size from about 10 microns (a micron is 10^{-6} meter) to a centimeter or so for large rain drops and snowflakes. Particles smaller than about 0.1 mm fall slowly because their motion is strongly limited by drag, so to a first approximation they can be considered to move with the air like a gas. Particles that fall more quickly are said to "precipitate"; their fall speeds are close to the "terminal velocity" at which their weight is balanced by aerodynamic drag. The drag increases with the density of the air, and with the square of the fall speed. Large rain drops fall at about 5 m s^{-1}, relative to the air. If such drops find themselves in an updraft with a speed faster than 5 m s^{-1}, they will actually be carried upward by the air.

The liquid and ice particles that make up clouds are typically "nucleated" on aerosols; the variable abundance of these cloud-nucleating particles, which are called cloud condensation nuclei (CCN), is a factor influencing the formation, number, and size of cloud particles. The number of CCN does not strongly influence the probability of cloud formation or the area-averaged rate of precipitation, under realistic conditions. The fascinating complexities of clouds are discussed further in later chapters.

Clouds and aerosols make up a tiny fraction of the atmosphere's mass, but they powerfully affect the flow of electromagnetic radiation through the atmosphere. This is also discussed further in later chapters.

LAYERS

There is no well-defined "top of the atmosphere" because the density of the air just continues to decrease exponentially toward zero at great heights. Nevertheless, all but about one millionth of the mass of the atmosphere lies below a height of 100 km, so it is reasonable (and somewhat conventional) to use this height as the atmosphere's top. Although 100 km sounds like a pretty thick layer of air, the radius of the Earth is a much larger 6,400 km. Relative to the Earth as a whole, the atmosphere is almost like a coat of paint, as can be seen in Figure 1.2. Life on Earth critically depends on that thin, fragile, mostly transparent, wispy shell.

The atmosphere is conventionally divided into layers. The lowest layer is the *troposphere*, which is in direct contact with the *lower boundary*, a term used by atmospheric scientists to refer to the top of the ocean and land surface. As discussed later, the lower boundary can be viewed as an energy source for the atmosphere. Roughly speaking, the atmosphere is "heated" by contact with the boundary;

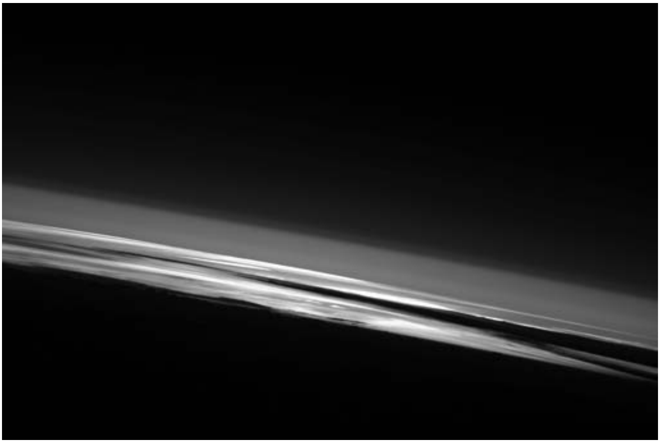

Figure 1.2. A photograph of the atmosphere as seen from space; the Sun is just below the horizon.

If the figure were in color, you would see that the lower atmosphere looks red because the blue photons emitted by the Sun are scattered away from the line of sight. A few clouds can be seen as dark blobs, blocking the Sun's rays.

Source: http://upload.wikimedia.org/wikipedia/commons/1/13/sunset_from_the_ISS.JPG

the heat enters directly into the base of the troposphere. In response, the troposphere churns like a pot of water on a stove, as buoyant chunks of air break away from the lower boundary and float upward, carrying energy (and other things) with them. The upper-level air is cooled by emitting infrared radiation to space. Once cooled, it sinks back to near the surface, where it is heated again, and the cycle repeats, in an irregular and chaotic fashion. The buoyancy-driven or "convective" motions act like a gigantic heat pump, cooling near the surface and warming above.

You may be wondering, if buoyancy pushes warm air up and cold air down, then shouldn't we find the warm air on top and the cold air below? Why does the temperature typically decrease upward? A partial explanation is that sinking air is warmed by compression as it moves to greater pressures near the surface and rising air is cooled by expansion as it moves to smaller pressures aloft. Further discussion is given in Chapter 3.

The small- and large-scale convective motions that are driven by surface heating and upper-level cooling extend through the lowest 10 or even 20 km of the atmosphere, depending on place and season. This layer of active weather is the troposphere. It contains about 80% of the atmosphere's mass.

The lowest portion of the troposphere is often called the *boundary layer* because it is strongly affected by direct contact with the lower boundary, that is, the ocean or the land surface. The boundary layer is turbulent; in fact, this is a key part of its definition.

The top of the troposphere is called the "tropopause." You have probably been there. Commercial airliners typically cruise near the tropopause level, at a height of about 12 km and a pressure of about 20,000 Pa.

You may feel a bit surprised that the troposphere has a well-defined top. As discussed in a later chapter, active weather, which is a characteristic property of the troposphere, is a kind of turbulence. Turbulent masses of air often have well-defined boundaries because turbulence tends to grow by active annexation of the surrounding, quiet fluid, as if by an advancing army.

Above the tropopause is the stratosphere. It is meteorologically quiet, compared to the troposphere. A distinctive property of the stratosphere is that the temperature generally increases upward there, as can be seen in Figure 1.1. The upward temperature increase is caused by heating of the middle and upper stratosphere due to the absorption of UV radiation from the Sun. The UV is actually absorbed by ozone (O_3), which is thereby converted into molecular oxygen (O_2) and atomic oxygen (O). The O_2 and O then recombine (in the presence of other species) to create another ozone molecule, so that there is no net loss of ozone. The net effect is a continual destruction and regeneration of ozone, accompanied by a heating of the air. The ozone cycle is an essential characteristic of the stratosphere.

The absorption of solar UV by ozone is important for life because UV causes biological damage if and when it reaches the biosphere. The actual amount of ozone in the atmosphere is tiny. Like water vapor, ozone is an atmospheric constituent that is *not* well mixed. In the middle stratosphere, about 30 km above the ground, the ozone concentration reaches a maximum, but it is a very small maximum: only about 1/100,000 of the air, depending on place and season. In the troposphere, ozone is much less abundant, except where it is artificially introduced as a pollutant.

The top of the stratosphere is called the stratopause. It is typically about 60 km above the surface. The troposphere and stratosphere together contain about 99.99% of the mass of the atmosphere.

It is conventional to define more layers above the stratopause, including the following:

The *mesosphere*, which extends from the stratopause to about 90 km, and within which the temperature decreases upward. The stratosphere and mesosphere together are often referred to as "the middle atmosphere"—no relation to Tolkien's Middle Earth. Reentering spacecraft heat up in the mesosphere, and most meteors burn up there. The mesopause coincides with what is sometimes called the "turbopause," the level above which mixing by the winds is no longer dominant.

The *thermosphere*, from the mesopause to about 500 km. Here the atmospheric gases are no longer well mixed and fractionate by their molecular weights. Molecular viscosity is strong enough to suppress turbulent mixing. The air is ionized by UV radiation, and as a result its motions can be influenced by electromagnetic fields. The temperature increases upward and is sensitive to the level of solar activity. The International Space Station orbits within the thermosphere.

The *exosphere*, from 500 km to an indefinite height. Within the exosphere, the atmosphere consists of rarely colliding individual molecules, mostly hydrogen and helium. Some of the molecules attain escape velocity and leave the Earth behind.

In this book, we will focus mainly on the troposphere.

CHAPTER 1

THE WINDS

The wind is the motion of the air. Typical wind speeds range from a few meters per second or less near the surface, to 50 m s^{-1} in the jet streams near the tropopause level, to 100 m s^{-1} in a strong tornado. These air currents arise in response to forces acting on the air, including pressure forces, which are often produced by spatially varying patterns of heating and cooling. We say that the circulation of the atmosphere is "thermally driven."

The term "scale" has already been used several times in this book. Atmospheric processes often have characteristic space and/or time scales. The concept of scale is particularly useful in connection with the winds and ocean currents. Spatial and temporal scales tend to increase or decrease together. Turbulent eddies are meters to hundreds of meters across and last for seconds to minutes. Thunderstorms are a few kilometers across and last for an hour or two. Winter storms can be thousands of kilometers across and last for several days. Monsoons span thousands of kilometers and persist for months. These diverse scales of motion can strongly interact with each other. Large-scale weather systems can excite small-scale turbulence. The turbulence, in turn, exerts drag forces and other influences on the large-scale weather. Such "scale interactions" are extremely important and will come up repeatedly throughout this book.

Because the atmosphere is a thin shell on the large spherical Earth, air can move much further horizontally than it can vertically. As a result, the large-scale motion of the air is mostly horizontal, and typical horizontal wind speeds are generally hundreds or even thousands of times faster than typical vertical wind speeds. On smaller scales, however, the vertical winds are sometimes just as fast as the horizontal winds; this can be the case in thunderstorms, for example, which can contain narrow but intense updrafts and downdrafts with speeds of tens of meters per second—not a friendly environment for aerial navigation. The winds carry energy, moisture, and other things, from place to place. Even moderate wind speeds are fast enough to carry air between any two points on the Earth in a few weeks.

The atmosphere is a closed system; air moves around, but, with minor exceptions, it does not enter or leave the atmosphere. It is therefore natural to speak of atmospheric "circulations." Some of these circulations are closely associated with familiar weather phenomena and reoccur, with minor variations, often enough so that we find it useful to give them names. Examples are thunderstorms, fronts, jet streams, tropical cyclones, and monsoons.

One of the most important factors influencing the circulation is the rotation of the Earth, which has profound effects on large-scale circulations. Because of the Earth's rotation, a point on the Earth's surface is moving toward the east at a speed that depends on latitude and is fastest at the Equator. Here is a number that is easy to remember: At the Equator, a point on the Earth's surface is moving toward the East at about 1,000 miles per hour (about 450 m s^{-1}). The Earth's rotation promotes the formation of beautiful atmospheric vortices. The vortices of many sizes and varieties are not details; they are essential features of the winds. Most of the kinetic energy of the atmosphere is associated with vortices of one kind or another.

Mountain ranges exert a major influence on the circulation because the moving air must go over or around such obstacles. The major mountain ranges of the world cause prominent wavy patterns in the global circulation. Some of the waves span thousands of kilometers and strongly influence the weather. Mountains also influence the pattern of atmospheric heating and cooling.

Small-scale motions are often driven by buoyancy forces associated with clouds; these are the convective circulations mentioned earlier. Buoyancy-driven circulations occur on a wide range of spatial scales, from tens of meters to thousands of kilometers. The circulating air moves not only vertically but also horizontally, along paths that sometimes approximate closed loops. These circulations are the wind fields that are associated with what we call weather. Familiar examples of convective circulations include thunderstorms, which are comparable in width to a city; continent-sized winter storms in middle latitudes; and monsoons, which can be comparable in horizontal scale to the radius of the Earth.

LATENT HEAT

From a human perspective, precipitation is perhaps the single most important climate variable. Less obviously, cloud and precipitation processes are leading actors in the physical climate system and strongly influence the global circulation of the atmosphere. One reason for this is that water has an enormous latent heat of vaporization, which can be defined as the energy required to evaporate a given mass of liquid. The latent heat of water is about 2.5 MJ kg^{-1}. Here is an amazing fact: the 2.5 MJ needed to evaporate one kilogram of water is equivalent to the kinetic energy of one kilogram of mass (which could be the same kilogram of water) moving at about 2,200 meters per second, or 7 times the speed of sound! For comparison, the latent heat of nitrogen, the most abundant atmospheric constituent, is only about 0.20 MJ kg^{-1}, less than one-twelfth that of water.[11]

When you climb out of a swimming pool, water evaporates from your skin. The energy needed to evaporate the water comes from your skin, and as a result the evaporation makes you feel cold. That same 2.5 MJ kg^{-1} of latent heat is "released" when condensation occurs. You may have experienced latent heat release in a sauna. Liquid water that is tossed onto the hot coals of a sauna flashes into the air as vapor and then condenses onto your relatively cool skin. The latent heat release that accompanies this condensation makes your skin feel hot.

The gigantic latent heat of water is one of the reasons that "moist processes" (atmospheric science jargon for processes involving phase changes of water) are very important for the Earth's climate. This will be discussed in a later chapter.

CHAPTER 1

A FIRST LOOK AT THE ENERGY CYCLE

Good advice for students of the climate system is "Follow the energy." That is the overarching theme of the next three chapters of this book. A broad overview of the energy flow is given in Figure 1.3. The climate system absorbs energy in the form of sunlight, mainly in the tropics and mostly at the Earth's surface. Clouds, ice, and bright soils modulate this solar absorption by reflecting some of the sunlight back to space. A substantial portion of the absorbed solar energy is used to evaporate water from the

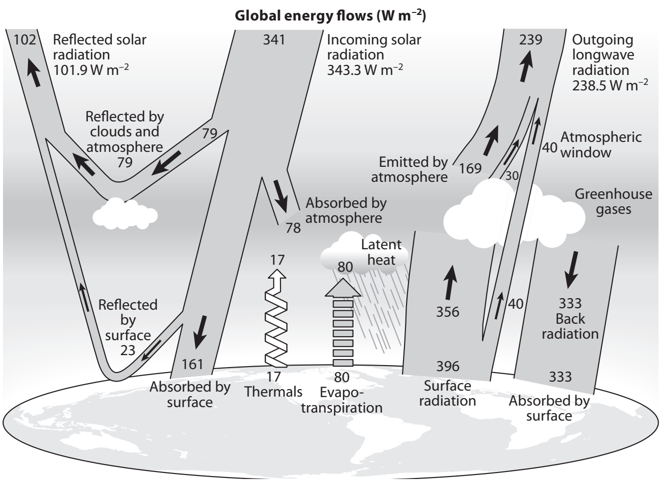

Figure 1.3. An overview of the flow of energy in the climate system.

The global annual mean Earth's energy budget for the March 2000 to May 2004 period (W m^{-2}). The broad arrows indicate the schematic flow of energy in proportion to their importance.

Source: Based on a figure in Trenberth et al. (2009).

tropical oceans. A larger portion enters the atmosphere in the form of infrared radiation emitted by the surface. The gases and clouds of the atmosphere absorb some of this, and reradiate the energy, again as infrared, both downward to the surface and upward to space.

Overall, considering both solar and infrared radiation, the atmosphere is radiatively cooled. The radiative cooling is balanced by the latent heat released when the water evaporated from the ocean recondenses to form clouds. In this and other ways, the Earth's energy and water cycles are closely linked. Chapters 3, 5, and 6 explore these links in more detail.

Atmospheric processes convert a small portion of the thermodynamic energy into the kinetic energy of atmospheric motion. The winds and the ocean currents carry the energy around from place to place on the Earth, cooling some regions and warming others, something like the heating, ventilating, and air conditioning system in a building.

The air also has gravitational potential energy, usually shortened to "potential energy," which increases (per unit mass) with height above the surface. The kinetic and potential energies combined are sometimes called the mechanical energy.

Ultimately, the energy that was originally provided by the Sun flows back out to space in the form of infrared radiation, called the "outgoing longwave radiation," abbreviated as OLR. The atmosphere strongly absorbs the infrared radiation emitted by the lower boundary and reemits the energy at the colder temperatures found aloft. Measurements show that, on time scales of a year or more, the globally averaged infrared emission by the Earth balances the globally averaged solar absorption very closely, within a couple tenths of a percent. The imbalance is believed to be just at the limit of what can currently be measured.

Radiative processes are key at both the beginning and the end of the energy story. They deserve a chapter of their own.