## 2. Building Blocks of the Climate System

In this section we will explore fundamental physical properties of the earth system which help us to describe and understand how the climate system works. One property is distance from the sun. Progressing outward from the sun, Mercury and Venus are too hot and Mars is too cold for life like we have here. Earth is located in a fairly narrow range of distance from the sun, in a "sweet spot" where water can exist in its liquid, solid, and vapor form, fostering life. At a very small scale, the properties of the different elements and molecules themselves determine the composition of the earth system, heating and cooling in the climate system, and effects on the biosphere. Consideration of temperature, pressure, and density help us to understand how the atmosphere and ocean behave. The visible and infrared parts of the *electromagnetic spectrum* are responsible for the greenhouse effect. *Absorption* and *emission* of electromagnetic radiation determines the temperature of our planet. Differences in heating give rise to winds and currents, which transport quantities around in what we call the general circulation of the atmosphere and ocean.

The 5 km level in the atmosphere is an interesting altitude to consider. About half of the molecules in our atmosphere lie below 5 km altitude, so the atmosphere is quite thin compared to the radius of the earth (6367 km). Since 5 km lies vertically in the middle of weather systems, winds at that level can be used to estimate their direction of motion. Near this level, the pattern of continental-scale waves excited by heating and topography provides a way for weather in one part of the world to affect weather in other parts of the world. These long-distance dynamical relationships are known as *teleconnections*. In addition, 5 km is the average altitude for the emission of infrared photons to space.

The electronic properties of molecules help us to understand how they interact with different parts of the electromagnetic spectrum. We will use simple laws of physics to calculate the strength of the natural atmospheric greenhouse effect. This may be contrasted with an actual greenhouse, where the roof serves to keep heated air in the greenhouse. The anthropogenic greenhouse effect is the degree to which the natural atmospheric greenhouse effect is augmented by our increased emissions of radiatively-active trace gases.

### 2.1 Important Constituents

Consideration of the atomic structure of the elements provides insight into fundamental properties of the earth system. Metals have many protons and neutrons and like to share electrons, which makes them dense. Metals are very common in the asthenosphere and lithosphere, which underlie the ocean and atmosphere. Quartz ($SiO_2$) is a very common crustal mineral (Table 2.1). Ca is often found in calcium carbonate, $CaCO_3$, or limestone, which is a byproduct of life in the ocean. The diversity of minerals has increased along with diversity of life forms over geological time.

| Lithosphere | Ocean | Atmosphere |
|---|---|---|
| 47% O | 55% $Cl^-$ | 78% $N_2$ |
| 28% Si | 30% $Na^+$ | 21% $O_2$ |
| 8% Al | 8% $SO_4^=$ | 1% Ar |
| 5% Fe | 4% Mg | 0-4% $H_2O$ |
| 4% Ca | 1% K | 0.04% $CO_2$ |

**Table 2.1.** The five most common components in the lithosphere, ocean, and atmosphere.

Considering the periodic chart of the elements (not shown), noble gases in the right-hand column, such as Ar, have no unpaired electrons, so they do not react with other molecules and exist in significant quantities in the atmosphere. Electrons love to be paired up. An element in the second column from the right have compact, powerful nuclei and only needs one more electron, making it good at attracting weakly-held electrons from other molecules. With an extra electron it becomes a negatively charged ion. An element in the left-hand column has a single unpaired electron in its outer shell, which can be readily lost, which makes it a positively charged ion. As a result, $Na^+$ and $Cl^-$ and other ions are common in the ocean and in life forms (Table 2.1).

The elements O, N, and C like to share multiple electrons and, together with H, make strong bonds that are useful for storing and manipulating energy in living organisms. In the atmosphere, the molecules $N_2$ and $O_2$ are unreactive at typical earth temperatures and are quite abundant (Table 2.1). Shortly we will examine the question of how relative abundances in the atmosphere evolved over geological time.

Special nutrients P, K, and N are important for life. The electronic properties of metals enable essential metabolic processes such as respiration (Fe) and photosynthesis (Cu). Others, such as Hg, Cd, Pb, and Al, can interfere with normal biomolecular functioning for both animals and plants.

With increasing anthropogenic $CO_2$ being emitted into the atmosphere, more $CO_2$ is dissolving into the oceans. When $CO_2$ dissolves it creates carbonic acid ($HCO_3$), which increases the acidity of the ocean. *Diatoms* are able to make use of the dissolved silicon and oxygen in the ocean to make their shells out of opal ($SiO_2$). They dominate the base of the food chain in colder oceans. *Coccolithophores* make their shells out of calcium carbonate ($CaCO_3$) and dominate the base of the food chain in midlatitudes. Increased acidity can shift the balance between major ecosystems in the ocean by tending to dissolve calcium carbonate shells, allowing the diatom ecosystem to expand at the expense of the coccolithophores and all of the life forms that they support.

In the atmosphere, the most common gas after $N_2$ and $O_2$ is $H_2O$, which varies from 4% in the humid tropics to almost 0% in coldest regions (Table 2.1). Since $H_2O$ is a polar molecule, it has a rotational dipole, with a plus charge near each hydrogen nucleus and a minus charge near the stronger oxygen nucleus, which is able to attract the electron cloud nearer to it. At earth temperatures, all molecules rotate and vibrate. Molecules with an electric dipole are special because they have an oscillating electric field, with quantized energy states. This enables such molecules to readily absorb and emit quantized packets of energy corresponding to specific wavelengths of infrared radiation.

Quantization of emitted and absorbed electromagnetic energy is evident in observed line spectra, where the wavelength of light corresponds to specific energy amounts. More complex molecules such as $N_2O$ or CFCs have more vibration/rotation modes, more absorption lines and more wavelengths, and are therefore more potent greenhouse gas molecules, molecule by molecule. However, since water vapor molecules are so numerous in the atmosphere, it is the most important greenhouse gas. Clouds and aerosols also absorb and emit infrared efficiently. Together with water vapor, they help keep the planet comfortable for life on earth. $N_2$, $O_2$, $H_2$, and the noble gases are symmetric and lack a rotational dipole, hence they are bad at absorbing and emitting infrared radiation and make very poor greenhouse gases.

Table 2.2 highlights some of the primary chemical actors on the stage of climate change. The first group includes five relatively long-lived compounds which are relevant to greenhouse warming and ozone depletion. The second group of five is important in air pollution problems.

| Constituent | Formula | Concentration | Source | Lifetime |
|---|---|---|---|---|
| carbon dioxide | $CO_2$ | 420 ppmv | combustion | 4 years |
| methane | $CH_4$ | 1.9 ppmv | land use | 10 years |
| nitrous oxide | $N_2O$ | 0.33 ppmv | fertilizer | 170 years |
| chlorofluorocarbons | CFCs | 3 ppbv | industrial | 80 years |
| stratospheric ozone | $O_3$ | 1-10 ppmv | sunlight | weeks |
| tropospheric ozone | $O_3$ | 0-200 ppbv | combustion | hours |
| carbon monoxide | CO | 0-100 ppbv | combustion | months |
| odd nitrogen | $NO_x$ | 0-50 ppbv | combustion | days |
| odd sulfur | $SO_x$ | 0-50 ppbv | combustion | days |
| hydroxyl radical | $OH^-$ | $10^{-11}$ | | $10^{-3}$ sec |

**Table 2.2.** The major climatically important trace gases in the atmosphere, with their concentration, source, and lifetime. The chemical lifetime of a constituent is defined to be the amount of that constituent in the atmosphere divided by its loss rate. For $CO_2$, the loss rate considered is due to photosynthesis.

The long-lived greenhouse gases $CO_2$, $CH_4$, $N_2O$, and CFCs have many vibration-rotation modes and are therefore excellent at absorbing and emitting infrared radiation. CFCs also reduce stratospheric ozone, letting more harmful UV reach the surface. In polluted air, members of the second group directly affect human health. Tropospheric ozone exceeding 65 ppbv is increasingly bad for human lung tissue and is associated with significant mortality. Odd nitrogen ($NO_x = NO + NO_2$) and odd sulfur ($SO_x = SO + SO_2$) form nitric acid ($HNO_3$) and sulfuric acid ($H_2SO_4$), which readily dissolve into hydrometeors. These fall out as acidic precipitation, harming ecosystems and human creations. $NO_x$ and $SO_x$ tend to rain out within a week or so. Considering typical wind speeds, acid rain tends to spread downstream of sources over continental scales. The hydroxyl radical (OH) plays an important role in partitioning species within chemical families. $OH^-$ is essential for converting source pollution gases such as $NO_2$ and $SO_2$ into precipitable dissolved acids, and is therefore known as the "*tropospheric cleanser*". Increasing emissions of CO and $CH_4$ can reduce $OH^-$ and thereby hamper its ability to cleanse the atmosphere of other pollutants.

*Focus: A Breath of Air*

If you release dye into the air, it will quickly mix with its surroundings. Turbulent motions, at scales ranging from centimeters to thousands of kilometers, transport air molecules and small particles around and mix them with their surroundings. Within a year, long-lived trace constituents are well-mixed throughout the troposphere.

An application of this principle to a single breath of air vividly emphasizes the interconnectivity of the earth system. Take in a breath. It is approximately 1 liter of air. At sea level near room temperature, 22 liters of air contain Avogadro's number of molecules: $6 x 10^{23}$ molecules. This means that your breath contains ~$3 x 10^{22}$ molecules! You may now exhale. After one year, the molecules from your breath will be spread evenly throughout the troposphere. How dilute will they be? To figure that out we need to know what the volume of the atmosphere is. If we could compress the atmosphere uniformly to sea level pressure, it would be about 8 km thick. The surface area of the earth is ~ $4 \pi (6 x 10^6 m)^2$, so the volume of the air at sea level pressure would be ~$3 x 10^{18} m^3$ or ~$3 x 10^{21}$ liters. If you divide the number of molecules in your breath by this volume there should be at least 10 molecules per liter! Each breath of air spread throughout the atmosphere would provide about 10 molecules for every liter sampled. Of course, your exhaled molecules may be incorporated into other parts of the earth system. You are inhaling molecules from all parts of the biosphere. Each breath contains molecules from every breath that you and every other organism exhaled last year.

We all share the same molecules eventually. Each of our exhalations literally becomes, molecule by molecule, one with the earth system, while each breath literally merges molecules from everything else into ourselves. It is hard to escape this molecular reality of the interconnectedness between large and small. Heroes and villains, plants and animals, all have shared some of the same molecules with you. By taking in air, water, and food, and metabolizing many of the molecules, our bodies are thought to replace approximately half of our molecules each year. It is sort of amazing that our appearance doesn't change very much. This steady interchange of molecules illustrates one facet of interconnectedness in the web of life.

## 2.2. Temperature

One important descriptor of climate is temperature, which is related to how fast molecules are moving, which depends on the amount of internal energy. At room temperatures, molecules travel at an average speed of 450 m/s in random directions. (Coincidentally, this is similar to the eastward speed of the earth's surface at the equator.) The Fahrenheit and Celsius (or Centigrade) temperature scales are based on phase change of water substance, while the Kelvin temperature scale is based on the concept that absolute zero represents a state of zero internal energy. This zero reference point of the Kelvin temperature scale is an essential feature for successful application in all scientific calculations. The Kelvin enjoys full honor alongside the meter, the second, and the kilogram as a fundamental unit in the System Internationale (SI).

Some insight into the Fahrenheit scale may be had by driving in the wintertime in Wisconsin, where they salt the roads. If it is below 0°F on your way to work, then there is usually a powdery white film on the roads. If it is above 0°F on your way home, then it is probably glistening wet. 0°F is defined to be the coldest temperature that a mixture of salt, ice, and water can be. The freezing point is 32°F for unsalted water, with180 gradations or degrees from freezing to the boiling point at 212°F. The Centigrade scale employs 100 degrees between the freezing point of water (0°C) and boiling point (100°C). You may convert back and forth by using $°C = 5/9(°F - 32)$ and $°F = 9/5\ °C + 32$. The Kelvin scale retains the 100 degrees between freezing and boiling, but starting from absolute zero, freezing doesn't occur until 273 K, so $K = °C + 273$.

Pure water will not freeze until 233 K unless it comes in contact with a foreign particle or surface capable of initiating condensation. (Coincidentally, this occurs where the Fahrenheit and Centigrade scales cross at -40°F = -40°C.) This means that a mud puddle won't freeze at 32°F without the mud particles. This also means that snowflakes and raindrops cannot form in the air above -40°F without smoke, dust, or other aerosols to act as nuclei [e.g., Wallace and Hobbs, 1977]. Ultimately, forest fires and windblown soil enable precipitation to fall from the sky.

How does the temperature of the earth compare with Venus and Mars? The earth's average surface temperature is 288 K (15°C, 59°F), about 33 K warmer than it would be without an atmosphere. The surface temperature of Venus is ~750 K, while that of Mars is ~215 K. Venus suffers from a runaway greenhouse effect due to 90 earth atmospheres of $CO_2$. The atmosphere of Mars is less than 1% of the earth's atmosphere. Its temperature is almost the same as if it had no atmosphere at all. To better understand these striking differences, we need to consider the transfer of electromagnetic energy.

**2.3. The Electromagnetic Spectrum**

In order to understand ozone depletion and the atmospheric greenhouse effect we need to consider the range of wavelengths emitted by matter in the universe, the spectrum of electromagnetic radiation, or light. Light is envisioned to have a wave-particle nature, with a photon having a specific wavelength characterized by an oscillation between electric and magnetic fields. The oscillation's wavelength is precisely related to the discrete amount of energy associated with the photon. Shorter wavelengths correspond to more energetic photons. The precise amount of energy in a photon corresponds to the amount of energy that was lost by the emitting molecule, whose vibration, rotation, and electronic states are quantized. Once emitted, light travels at $3 \times 10^8$ m/s, until absorbed by another molecule, which can only absorb the photon if it can increase its energy level by the exact quantum of energy corresponding to its wavelength. Therefore, the ability of a molecule to absorb and emit electromagnetic radiation depends on its unique structure and energy states. Indeed, the spectral absorption features of every molecule provide a unique "fingerprint" for identifying their presence, which is useful in monitoring concentrations. Knowledge of the absorption spectra for different gases is crucial for calculations of heating rates in the atmosphere.

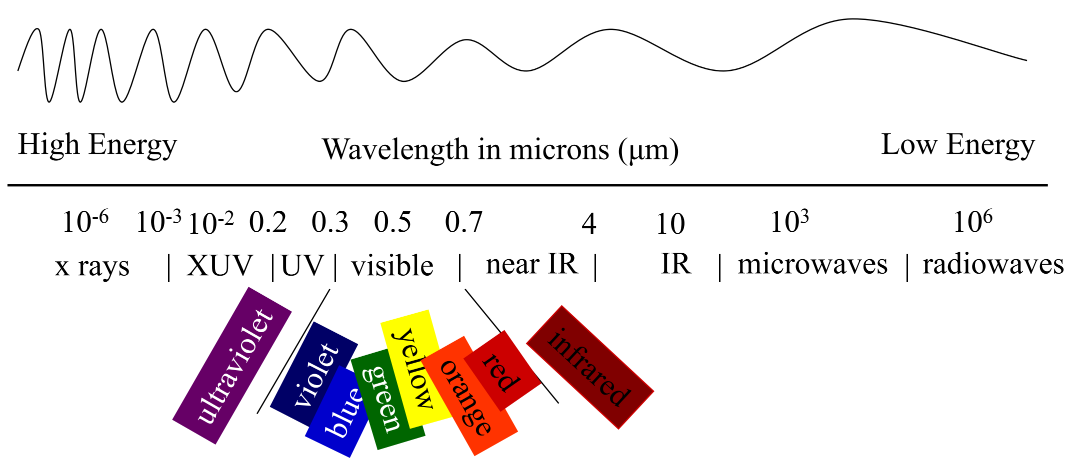


**Figure 2.1.** The electromagnetic spectrum. The energy of a photon is inversely proportional to its wavelength.

Figure 2.1 shows how we categorize different parts of the electromagnetic spectrum by wavelength, ranging from very short, energetic x-rays to very long, mellow radio waves. The wavelength of light is often given in microns, or millionths of a meter. The most commonly emitted photon from the sun, yellow light, is about 0.5 microns (0.5 $\mu$m), or $0.5x10^6$ m wavelength. 1 micron is about 1/1000th the width of a pencil stroke. One micrometer = 1 $\mu$m = $10^{-6}$ m = 1 micron = $10^3$ nanometers = $10^4$ Angstroms. The diameter of $N_2$ and $O_2$ molecules is about 0.001 $\mu$m, so visible light doesn't interact strongly with most air molecules. We have evolved to see in the visible part of the spectrum because the photosphere of the sun, being 6000 K, emits most of its energy in the visible. The familiar spectrum of light visible to humans ranges from the shortest, violet (0.3 $\mu$m), to blue, green, yellow, orange, and the longest, red (0.7 $\mu$m). Plants present flowers which are attractive to humans in the visible, but also to insects which see in the UV (0.1-0.3 $\mu$m). XUV and x-rays are found at even shorter wavelengths. Toward longer wavelengths from red one encounters near-IR (0.7 - 4 $\mu$m), IR (4 - 100 $\mu$m), microwaves (cms), and radio waves (meters). Objects at earth temperature emit energy primarily in the IR near about 10 $\mu$m.

For human color vision, there are three photo-pigments in the cone cells of the eyes that exhibit three distinct light absorption curves sensitive to violet, green-yellow, and yellow [Mollon 1995]. About 6% of males have photopigments with green-yellow and yellow absorption curves closer together, while another 2% of males lack the green-yellow photo-pigment altogether. Thus, 1 out of 12 males see the world differently than the rest of humanity and the perception of color is rooted in specific molecules. Some scientists have speculated that these males may be better at perceiving shades of gray near twilight, with superior hunting having allowed them to succeed. Specific molecules are often at the root of irreducible sensory impressions, imparting a distinct noumenal essence. Other examples will be discussed in Chapter 10.

Figure 2.2 focuses on the middle part of the spectrum in Figure 2.1, from 0.1 to 100 $\mu$m, which includes most of the solar and infrared energy emitted from the sun and the earth. Compare the spectrum of electromagnetic energy emitted from the sun (at 6000 K) to that emitted from the earth (at 255 K). For both the sun and the earth, the power output depends on wavelength. The average wavelength of emission is much shorter for the sun than for the earth. There is a clear separation near 4 $\mu$m. *Shortwave radiation* is solar radiation, which peaks in the visible part of the spectrum, being characterized by wavelengths near 0.5 $\mu$m, or yellow. *Longwave radiation* is earth radiation, which peaks in the infrared part of the spectrum, being characterized by wavelengths near 10 $\mu$m. This separation is useful in discussing the greenhouse effect. Atmospheric molecules interact quite differently with visible and infrared radiation. Air is nearly transparent in the visible, letting the sun heat the surface, but some kinds of molecules absorb well in the infrared. It is the interplay of infrared and visible energy that constitutes the greenhouse effect.

The curves in Fig. 2.2 have been drawn to have the same area underneath them, to signify the balance of absorbed solar radiation, at the earth's distance from the sun, and emitted infrared radiation in determining the temperature of the earth. To show how much power is emitted from the surface of the sun compared to the surface of the earth, the curve on the left would have to be made some 300,000 times larger, completely dwarfing the curve for emission from the earth. The power of energy emitted from the photosphere of the sun is 70 million W $m^{-2}$ at 6000 K, while the power of energy emitted from the earth is 240 W $m^{-2}$ at 255 K. The power of emitted energy increases very rapidly with increasing temperature, being 390 W m$^{-2}$ for emission at 288 K. This relationship is captured by the Stefan-Boltzmann Law,

$$E = \sigma T^4 , \qquad (2.1)$$

where $\sigma$ = 5.67x10$^{-8}$ W m$^{-2}$ K$^{-4}$ is the Stefan-Boltzmann constant, $T$ is in Kelvins, and $E$ is the power of emitted radiation in W/m$^2$. This very strong dependence of emission on temperature is an essential brake in the climate system which keeps temperatures from becoming extreme, a topic that we will return to shortly. It constitutes a strong *negative feedback* to a temperature change.

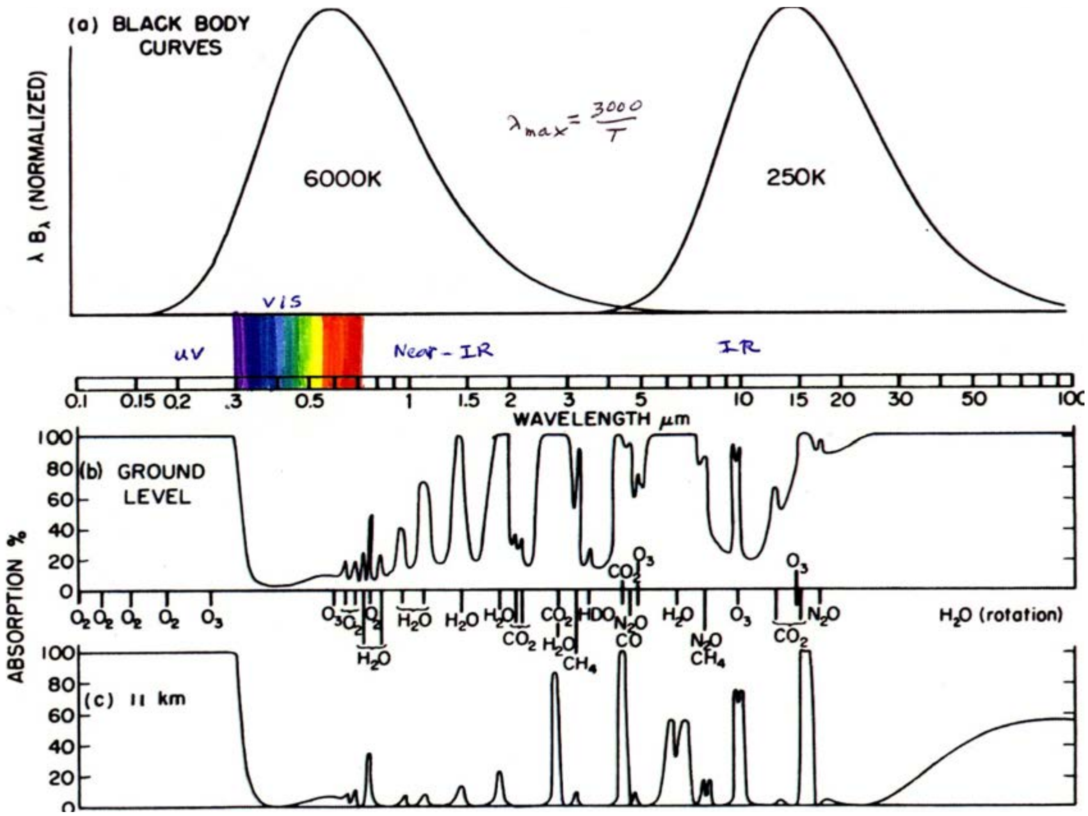


**Figure 2.2.** a) The solar and earth emission spectra, b) percent absorption for the whole atmosphere, and c) percent absorption for the part of the atmosphere that lies above 11 km.

About 2% of the solar energy is in the form of ultraviolet light (UV, 0.2-0.3 $\mu$m). In the rarefied stratosphere, this small amount of total energy that is absorbed by ozone causes temperatures to increase upward. UV photons are individually far more energetic than visible or infrared photons, which is why they are dangerous to biomolecular processes. It is good that ozone in the stratosphere absorbs most of the UV, protecting life on earth. Since less than 2% of the solar spectrum is in the UV, a slight increase in UV due to reduced stratospheric ozone is not the primary cause of anthropogenic greenhouse warming, which is due to increased trapping of infrared radiation.

Perhaps the most striking aspect of Fig. 2.2 is the difference between the average wavelength for solar versus earth emission. This inverse relationship between the temperature of the emitting body and the wavelength of emitted radiation makes intuitive sense. If you take a rope and shake it up and down quickly you will observe short waves moving away from you along the rope. If you do it less energetically, longer waves will move along the rope. Hotter molecules are more energetic and will emit more energetic photons of shorter wavelength. Cooler molecules vibrate less quickly and will emit less energetic photons of longer wavelength. This natural relationship between increasing temperature and decreasing wavelength of emitted radiation is expressed in Wien's Law:

$$\lambda_{max} = \frac{3000}{T} , \qquad (2.2)$$

where T is in Kelvins and $\lambda_{max}$, the wavelength of maximum energy emission, is in microns. From Wien's Law, we may easily compare the photosphere (6000 K) radiating at 0.5 m and the earth (300 K) radiating at 10 m.

We can get a hint about what it would be like to see into the infrared by turning on a stovetop element. As it warms up you can begin to feel the infrared being emitted from it, even though you can't see the electromagnetic radiation with your eyes. As it gets hotter, more total photons are emitted, and their average wavelength is getting shorter, with some photons being emitted in the deepest red that you can see. As it heats further, it may get hot enough for $\lambda_{max}$ to approach yellow. Then you'd better turn it off! Incandescent light bulbs heat air to 6000 K to glow like the sun in the visible. If the earth heated up, the emission spectrum in Fig. 2.2 would grow and the peak would shift to the left, toward shorter wavelengths. Wien's Law also shows that hotter stars glow blue, while cooler stars glow red.

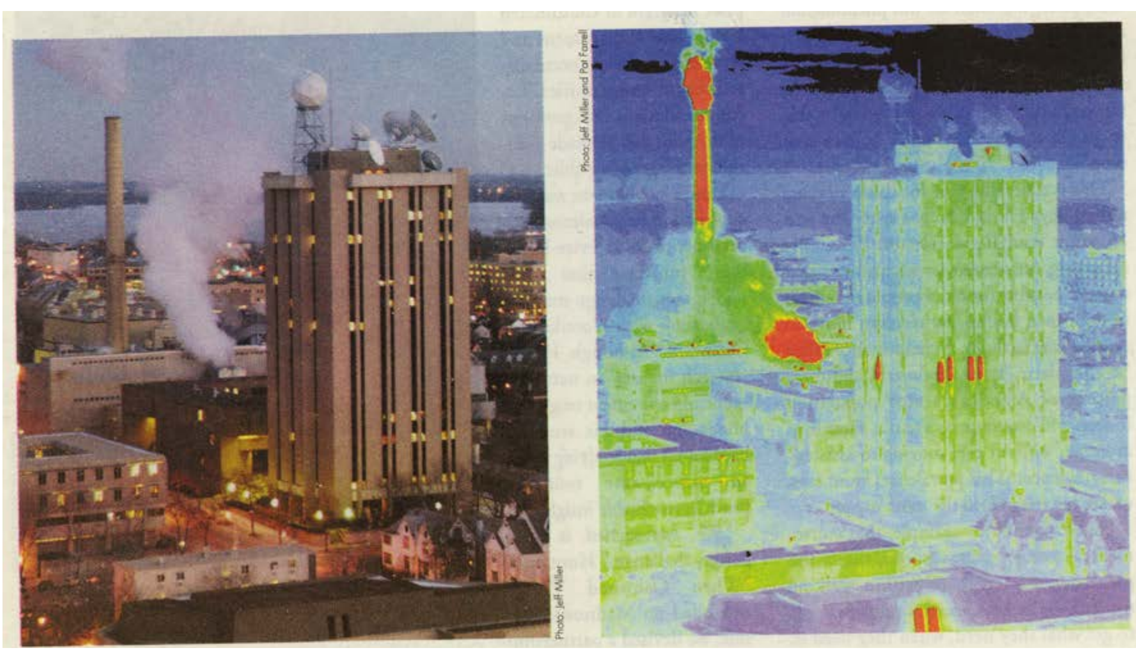


**Figure 2.3.** The Atmospheric, Oceanic, and Space Sciences building in Madison, WI on a cold winter day as seen in the visible (left) and color-coded infrared (right).

Figure 2.3 contrasts the appearance of the Atmospheric, Oceanic, and Space Sciences building in Madison, WI, as seen in the visible and infrared parts of the spectrum. The infrared signal is related to the number of infrared photons, which is related to the Stefan-Boltzmann Law (2.1), but it has been color-coded with blue indicating objects that emit little infrared (cold) and red indicating objects that emit a lot more infrared (hot). This emotional color-coding scheme is actually opposite in sense to Wien's Law, where cooler stars are red and hotter stars are blue.

## 2.4. Vertical temperature structure

Figure 2.4 shows the globally averaged vertical temperature structure of the atmosphere and the portions of the solar spectrum which warm each region. Half of the solar radiation is in the visible part of the spectrum. The atmosphere is largely transparent to visible light, so sunlight warms the earth's surface. The atmosphere *transmits* visible light well. The troposphere, or turning sphere, is heated from below like a pot of boiling water. At any moment, there are some 15,000 thunderstorms attempting to transport heat upward, away from the surface. Rising air expands and cools. This upward temperature decrease pauses at the tropopause, which occurs at ~ 8-10 km altitude and ~ 240 K in the extratropics, and at ~ 18 km and ~190 K in the tropics (Fig. 2.4).

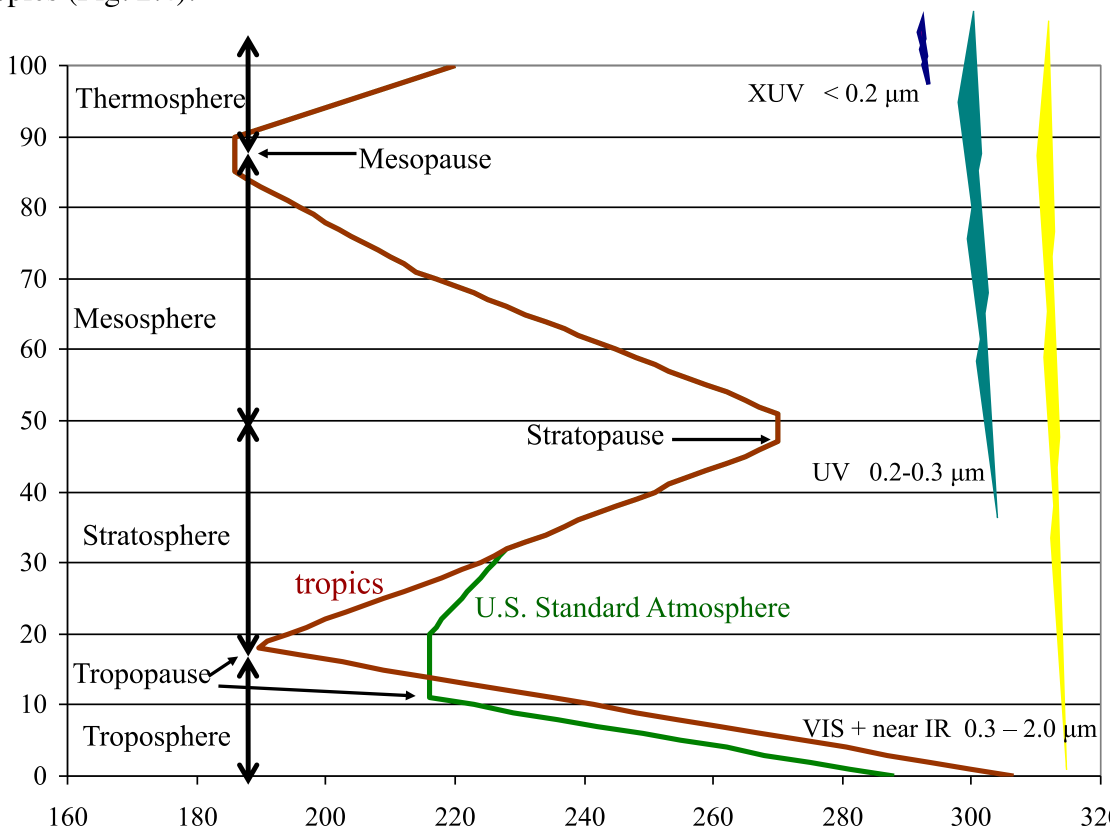


**Figure 2.4.** U.S. standard atmosphere (green) and typical tropical temperature profile (red).

In the stratosphere, ozone absorbs solar UV, causing temperature to increase upward. This region is very stable and characterized by stratified motions. It takes 5-10 years for air to be lofted upward through the stratosphere by solar heating. The upward increase in temperature pauses at the stratopause near 50 km altitude. In the middle region, or mesosphere, molecules radiate infrared to space but absorb very little energy from the sun. The coldest regions of the atmosphere are found at the mesopause near 80 km altitude. Temperature increases rapidly into the thermosphere, where the most intense photons are absorbed through photodissociation and photoionization. During solar maximum, the thermosphere is much hotter and expands outward, when XUV can be 2-3 times stronger. This expansion helps to clear lower earth orbits of space debris. The three warm layers occur due to the absorption of three different solar wavelength bands, with the most intense, shortest wavelength photons absorbed highest up: XUV in the thermosphere, UV in the stratosphere, and visible light and near-IR at the surface.

In the ocean surface layer, temperatures range from -4°C to 40°C. Below the sunlit upper few hundred meters the ocean is uniformly cold, in the narrow range -2°C to 4°C. If the ocean is heated from above, a stable layer is created, with lighter water near the surface, which inhibits heat exchange with the deep ocean. If the ocean is cooled from the surface it can get cold and dense enough to sink. If the surface gets too cold, it will form a thermally protective layer of sea ice, which floats on top. These processes keep the ocean fairly uniform in temperature.

## 2.5. Pressure and density

Most air molecules are about $10^{-9}$ m in diameter. At sea level pressure air molecules fly past each other about $10^{-8}$ m apart, travelling about $10^{-7}$ m before they collide. Near sea level, air molecules travel a distance of about 450 m in one second. But because they collide with about 5 billion other molecules in that one second, it is hard for a molecule to make it very far. In calm air molecules will travel about 1 cm in one day, the distance over which molecular diffusion is effective. These molecules all support each other against the downward pull of gravity through a tremendous juggling act. With each collision, a molecule changes its momentum, thereby exerting a force. Pressure is the force per unit area caused by molecules bouncing off of a surface. Your body is used to sea level pressure, exerting an equal and opposite outward pressure. If you were suddenly thrust into the near-vacuum of space, you would explode.

Pressure may be thought of as the weight of the molecules above you acting on each 1 $m^2$ surface area. Weight = mass x gravitational acceleration, which exerts a force. The pressure at a given altitude, P, is given by the mass of the molecules above altitude z, M(z), times gravitational acceleration, $g = 9.81$ m $s^{-2}$, divided by the surface area of the earth, $A_e = 5 \times 10^{14}$ $m^2$:

$$P(z) = \frac{M(z)g}{A}. \qquad (2.3)$$

As altitude increases, the mass of the molecules above that level decreases, so pressure decreases upward. The observed global mean sea level pressure is $10^5$ Nt $m^{-2}$. This can be used in (2.3) to determine the mass of the atmosphere: $10^5$ Pa x $[4\,\pi\,(6\times10^6\text{ m})^2]$ / 10 m $s^{-2} \sim 5 \times 10^{18}$ kg.

The SI unit for force is the Newton (N), while the SI unit for pressure is the Pascal (Pa), named after the French mathematician, Blaise Pascal. 1 Pa = 1 N $m^{-2}$. Sea level pressure may be described variously as 1 atmosphere, 1 bar, 1000 millibars, 1000 hPa (1 hPa = 100 Pa), or 100,000 Pa. Since $g \sim 10$ m $s^{-2}$, there are 10,000 kg of air pushing down on each 1 $m^2$. If you wanted to cross a newly frozen lake you would be less likely to break through the ice if you wore skis, because then your weight would be distributed over a larger area, reducing your pressure on potentially fragile spots on the ice.

*Focus: Hydrostatic Balance*

Have you ever wondered why gravity doesn't just crush all air molecules to the surface of the earth? Since each air molecule has mass and since gravity attracts each one toward the center of the earth, what keeps air suspended against gravity? Consider a parcel of air or a balloon (Fig. 2.5a). Since pressure decreases upward there are fewer molecules pressing from above than from below. This difference in pressure, or pressure gradient force, is directed upward and opposes the force due to gravity. Throughout most of the atmosphere and ocean these two forces balance nearly exactly, so the fluid is mostly static, *hydrostatic balance*. Imbalances between gravity and the vertical pressure gradient force lead to vertical accelerations, with updrafts and downdrafts in thunderstorms, and sinking plumes in the high latitude oceans. The fact that most of the atmosphere and ocean are in hydrostatic balance means that we can know the vertical pressure gradient force per unit mass to 3 significant digits. From hydrostatic balance,

$$-\frac{1}{\rho}\frac{\partial p}{\partial z} = g \,, \qquad (2.4)$$

the vertical pressure gradient force per unit mass can be accurately estimated to be ~ 9.81 m $s^{-2}$, the acceleration due to gravity. The vertical decrease in pressure exhibits an exponential shape, which can be understood by examining the interplay between pressure and density.

The downward rate of pressure increase is different in the atmosphere compared to the ocean (Fig. 2.5). This is because the atmosphere is much more compressible. Density, $\rho$, is the mass per unit volume (kg $m^{-3}$). If it is hard to change the density of a substance, such as water, we say that it is nearly incompressible. We know from using a bicycle pump that air is compressible. A flexible balloon ascending in the atmosphere will experience lower and lower pressure, so it will expand, with the same mass of air molecules occupying a larger volume, so the density decreases. Since the pressure is highest near sea level, air has the highest density there: (~ 1.2 kg $m^{-3}$). When you climb upward one meter near sea level you put many molecules below you, so pressure decreases rather quickly. Since pressure decreases upward quickly near sea level, so does density. At higher levels, where the air is less dense, travelling upward one meter will put fewer molecules below you, so pressure and density decrease upward more slowly. This interplay of pressure and density leads to *exponential profiles* of upward decrease in the atmosphere (Fig. 2.5a).

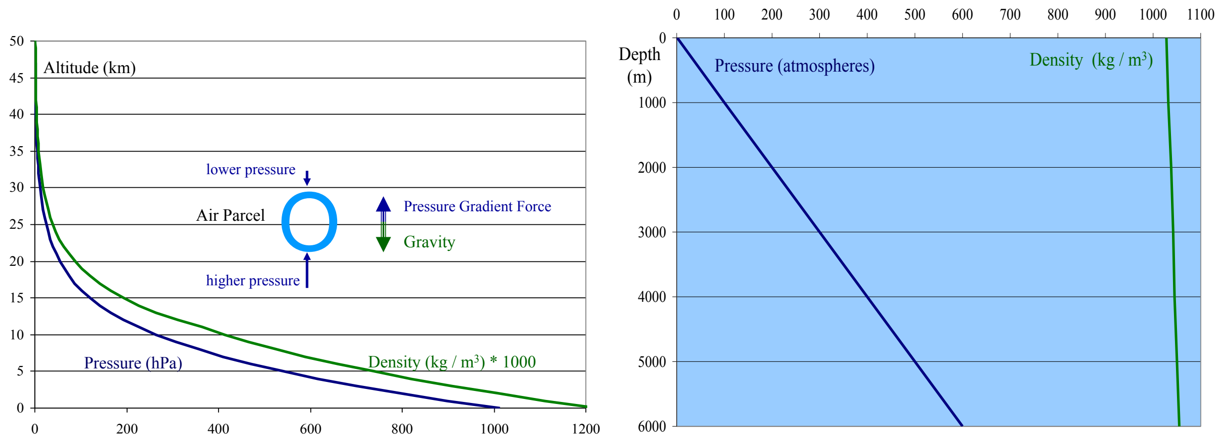


**Figure 2.5.** Vertical profiles of pressure (blue) and density (green) a) in the atmosphere and b) in the ocean, with illustration of hydrostatic balance.

Density and pressure increase linearly going down into the ocean (Fig. 2.5b), because water is nearly incompressible. Since molecules in a liquid are already touching each other, you can't cram them much closer together. The density of pure water is ~1000 kg m$^{-3}$ and that of sea water is ~1040 kg m$^{-3}$. Even at the bottom of the ocean, the density is only 5% greater than at the surface, due to pressure alone. The density of ice is variable, but averages about 10% less than water, so it floats on water, inhibiting the freezing of water from cold air above. The density of the crust is on the order of 3000 kg m$^{-3}$, so the oceans float on the lithosphere.

Since 1 m$^3$ of ocean water contains 1000 kg, every 10 m of depth adds another 10,000 kg of mass over each m$^2$, which, when multiplied by g, adds another 10$^5$ Pa or another atmosphere of pressure! Human divers under such intense pressures experience body volume reduction by up to 1/3. Below 200 m (20 atmospheres pressure) divers' eyes shrink, causing flashing auras. High pressure nervous syndrome can lead to body tremors, hallucinations, and death. At 1 km depth, there are 100 atmospheres of pressure. It is amazing that sperm whales can dive over 1000 m to grapple with the tasty, enormous squid known in Moby Dick as the kraken. At 5 km depth, the pressure is 500 atmospheres, rather an inhospitable condition for humans.

Surface pressure is a good measure of the climate-altering potential of the atmosphere. The surface pressure on Venus is about 90 earth atmospheres, while that on Mars is only 0.007 earth atmospheres. Although both atmospheres are mostly $CO_2$, the large amount on Venus, plus being closer to the sun, has led to a runaway greenhouse effect, while the minor atmosphere on Mars creates a negligible greenhouse effect. Sperm whales might survive the pressure at the surface of Venus, but not the heat at ~ 800 K.

## 2.6. Atmospheric Charts

Charts of the altitude of atmospheric pressure surfaces are extremely useful for diagnosing and predicting the weather and for understanding how various properties are transported within the atmosphere. *Radiosondes* are small weather balloons with instruments aboard that are launched simultaneously around the world, which radio information to ground stations whenever they reach specific pressure levels. From these simultaneous observations over the globe *synoptic charts* are made, showing winds, temperatures, and other properties at specific pressure levels in the atmosphere. Common chart levels include sea level, 850 hPa, 500 hPa, and 300 hPa. Figure 2.6 shows a NH wintertime average 500 hPa chart in polar stereographic projection. The contour labels show how far you would have to ascend from sea level to put half of the atmosphere below you. These contours are similar to contours on a topographic map, connecting lines of equal altitude above sea level.

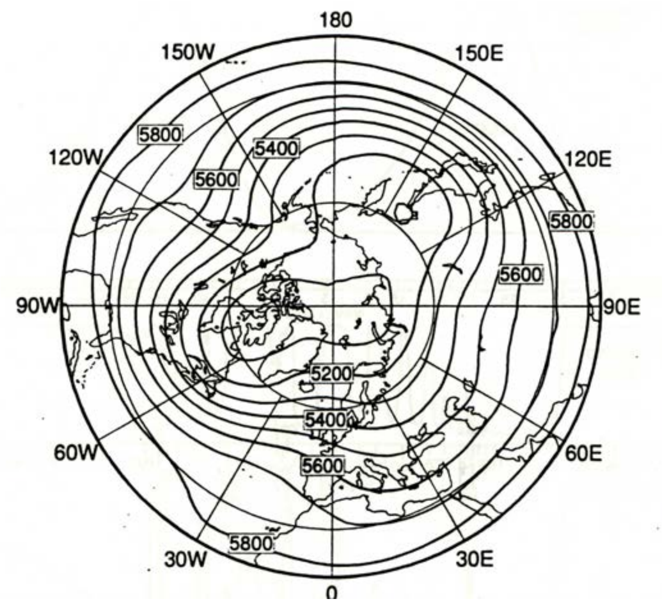


**Figure 2.6.** A Northern Hemisphere polar stereographic chart of 500 hPa geopotential height (in tens of meters) averaged for DJF, with contour interval 100 m. [Palmen and Newton 1969.]

Why are the contour numbers larger in the tropics and smaller over the pole? Warm air expands, as in a hot air balloon, while cold air contracts. In the tropics the warm air expands, becoming less dense, so that you have to go upward relatively far (~ 5800 m) to reach the 0.5 atmosphere level. In the cold dense air over the winter pole you only have to go up to ~ 5100 m altitude to put half of the molecules below you. Where warm air extends poleward, high height contours bulge poleward, indicating a *ridge*. Where cold air extends equatorward, low height contours bulge equatorward, indicating a *trough*.

Winds tend to blow parallel to height contours, counterclockwise around the north pole, from west to east. These are called the midlatitude westerly winds. The wind speed is fastest where lines are pinched together, similar to water flowing faster in a constriction. The jet stream is particularly strong near Japan and Eastern North America, due to the strong temperature contrast between the warm ocean and cold continent, making steeper height gradients. These winds control where weather comes from. We look toward the west for most of our weather. If you see a "red sky at night" it is a "sailors' delight", since there are no storm clouds low on the horizon, coming from the west, which would have obscured the sunlight.

The half-atmosphere level may be called the "steering level". Winds at 500 hPa are reliable indicators of where a severe thunderstorm will move, where high and low pressure systems come from, and how tropical cyclones recurve. These winds that blow nearly parallel to height contours are called *geostrophic* winds, named after the earth (geos) turning (trophic) [e.g., Holton and Hakim 2012]. It results from a near-balance between the tendency for air parcels to slide downhill toward the pole, but never getting there because the earth is turning around its axis. This bizarre result is due to the Coriolis effect, which we will study shortly. An air parcel in the midlatitudes has eastward momentum associated with the rotation of the earth. The horizontal pressure gradient force will cause the parcel to accelerate radially toward the rotation axis, so that it stays blowing parallel to height contours (Fig. 2.6). Figure 2.6 shows that, climatologically, there is usually a ridge over the west coast of North America, which steer storms which develop in the Pacific toward Alaska. Over cold eastern North America, winds approaching the trough bring small snowstorms from the northwest known as Alberta Clippers.

These ridges and troughs are part of patterns called planetary scale Rossby waves. A change in the planetary wave pattern implies mutual changes over the hemisphere. The 500 hPa level is useful for investigating how one part of the atmosphere is connected to another. One powerful variety of teleconnection is that energy emanating from regions of chronic tropical thunderstorms can modulate the planetary wave pattern in the extratropics. In some winters the ridge/trough pattern over North America is weaker or stronger than in others. Figure 2.7 shows changes in the 500 hPa height pattern during an El Nino event. Note the emanation of a planetary wave train out of the tropics. When the central Pacific is unusually warm and cloudy, this wavy pattern is disturbed, favoring more rain in California and in the southeastern U.S. This is one consequence of a natural climate variation in the tropical ocean and atmosphere that we will investigate called the El Nino Southern Oscillation (ENSO).

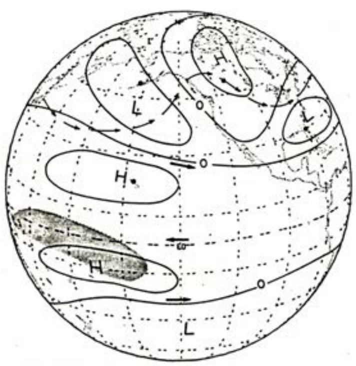

**Figure 2.7.** Schematic illustration of the hypothesized global pattern of middle and upper tropospheric geopotential height anomalies (solid lines) excited by El Nino during boreal winter. Shading indicates enhanced rainfall. Note the strengthening of the subtropical jets and ridging over Western Canada during El Nino. [Fig. 11 of Horel and Wallace 1981].

## 2.7. Radiative balance for the earth

We seek to understand the mechanisms of past and future climate change. Factors that control our climate include the intensity of the sun, how the earth orbits the sun, the distribution of continents, ice sheets and vegetation, volcanic eruptions, gaseous composition, the type and distribution of clouds and aerosols, the general circulation of the atmosphere and ocean, and other nonlinear aspects of our chaotic and deterministic climate system. We want to be able to make realistic estimates of the greenhouse effect and its likely change due to human activities. To approach this problem let us pose the first order question: Considering transfer of electromagnetic energy to and from the earth, what should its temperature be? We will calculate this *radiative equilibrium temperature*, $T_{re}$, using five well-understood physical principles: 1) energy absorbed equals energy emitted, 2) energy emitted depends very strongly on temperature, 3) energy absorbed depends on distance from the sun and 4) on planetary albedo, and 5) the emitting surface area is four times larger than the absorbing cross-section. Application of equation 1) to find $T_{re}$ is a powerful example of scientific reasoning.

*1) Radiative equilibrium.* In one full trip around the sun, the change in the globally-averaged surface temperature has been usually much less than $\pm$ 1 K in the historical record. During one year our planet receives and emits on the order of $10^{24}$ J. To a very good approximation, there is a balance between incoming and outgoing energy. This balance has to be very nearly exact, or the temperature would drift much more than 1 K in 1 year. At any location in the earth system, an exact state of constant radiative equilibrium does not occur. (Consider the rich array of temperature variations that exist in the earth system, including diurnal cycles, seasonal cycles, the equator to pole temperature gradient, and land-sea heating contrasts. ) Nevertheless, on a global and annual average, radiative equilibrium is an excellent approximation, which allow us to solve for the radiative equilibrium temperature $T_{re}$ from the equation

$$Energy\ In = Energy\ Out\ . \qquad (2.5)$$

Energy In is solar radiation, largely in the visible part of the electromagnetic spectrum, while Energy Out is mostly in the infrared part of the spectrum.

*2) Energy Out*. Theoretical and experimental physics has given us a precise formula for calculating the emission of electromagnetic radiation from a body at a given temperature, the Stefan-Boltzmann Law (2.2). The amount of energy in Joules (J) passing through 1 $m^2$ in 1 second is measured in W $m^{-2}$, where 1 W = 1 J/s. The strong temperature dependence of this emission provides a strong *negative feedback* in the climate system. Hot objects emit way more radiation than cold ones. If something acts to heat an object, it will emit much more radiation and cool itself off. If something acts to cool an object, it will emit much less radiation than it did before, allowing Energy In to warm it back up. If the temperature of an object increased to 2 times its previous value, it would emit $2^4$ or 16 times as much radiation! If the temperature decreased to 1/2 of its previous value, it would emit $(1/2)^4$ or 1/16 times as much radiation. Thus, a negative feedback is a process that opposes the initial tendency. A few selected values demonstrate the strong dependence of emission on temperature. The atmosphere near 5 km altitude radiates 240 W $m^{-2}$ at 255 K, the surface emits ~ 390 W $m^{-2}$ at 288 K, and our skin emits ~ 500 W $m^{-2}$ at ~ 308 K. The solar photosphere emits $70\times10^6$ W $m^{-2}$ at 6000 K. It is a good thing that we are far from the surface of the sun!

*3) The power of emitted radiation decreases by one over the distance squared*. Since the surface area of a sphere of radius r is $4\pi r^2$, the intensity of radiation is diluted by $r^{-2}$, as it passes outward through larger and larger spheres. The force of gravity, electrical attraction, and the power of sound also decrease by $r^{-2}$. The power of solar radiation, S, decreases by $r^{-2}$, where r is the distance from the sun. The power at distance $r_2$ is related to the power at the solar surface, $r_1$, by

$$S_2 = S_1 \left(\frac{r_1}{r_2}\right)^2 . \qquad (2.6)$$

Given a solar radius of $r_1$ ~ 660,000 km, emitted radiance $S_1$ ~ 70 x $10^6$ W $m^{-2}$, and the earth's distance to the sun $r_2$ ~ 150,000,000 km, the power of solar intensity at the earth's orbit is $S_2$ ~ 1370 W $m^{-2}$. This number is often called the "*solar constant*", although the intensity of solar radiation varies around the earth's elliptical orbit and also changes with the solar cycle. Sometimes there are no sunspots at all, such as during the Maunder minimum of 1615-1700. The earth is about 5 million km closer to the sun on January 4 than on July 4, making for 7% stronger solar power on January 4 than on July 4. The earth's ellipticity varies on a ~100,000-year time scale, which is one of three primary parameters in the orbital theory of climate change.

Figure 2.8 shows the variation of solar output as seen by a variety of instruments located outside of the earth's atmosphere. There are disagreements between instruments regarding absolute calibration of 1-6 W m$^{-2}$, which shows the uncertainty in our knowledge of the solar constant. However, the variation across the solar cycle is about 2 W m$^{-2}$ in all instruments, which shows that the range is fairly accurate. This range in output from solar maximum to solar minimum is about 0.1% of 1370 W m$^{-2}$. This variation of 2 W m$^{-2}$ is similar in magnitude to current estimates of increased downwelling infrared due to anthropogenic greenhouse forcing. The solar cycle is readily detected in changes in stratospheric temperatures.

To find out the solar power, $S_2$, near another planet at a distance $r_2$ from the sun, you can use (2.5) with $S_1$ = 1370 W m$^{-2}$ and $r_1$ = 150,000,000 km. We are almost done with estimating Energy In, but we need to remember that some of the sunlight is reflected and that a planet's shadow has the shape of a circle.

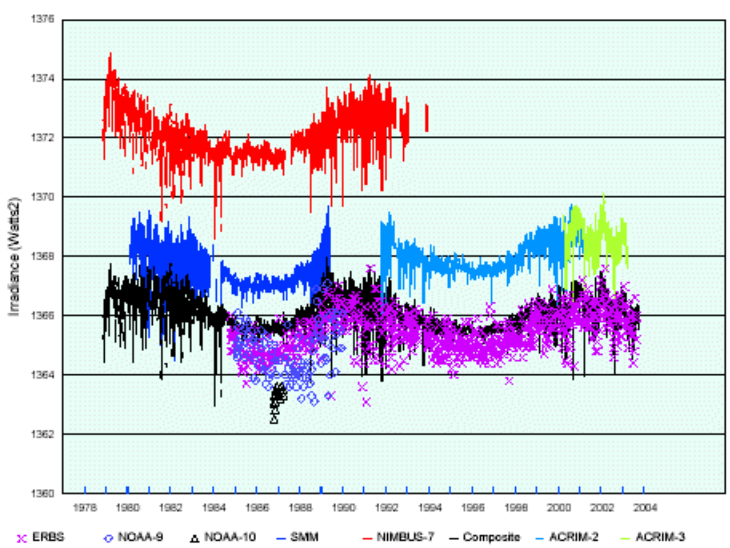


**Figure 2.8.** Measurements of solar power intensity made by a variety of satellites during 1979 – 2004 [http://www.ngdc.noaa.gov/stp/SOLAR/IRRADIANCE/irrad.html].

*4) Some surfaces reflect more sunlight than others*. The albedo, $A$, is the fraction of light reaching an object that is reflected. $A$ = 1.0 means that all incoming sunlight is reflected. Snow and ice surfaces have high albedos, ~ 0.8, because they reflect ~ 80% of the sunlight that strikes them. If more of the earth's surface is covered by ice and snow, then more sunlight will be reflected. With less energy absorbed, the earth would cool, making it easier for the ice and snow cover to grow. If some ice and snow were to melt, more energy would be absorbed, making the earth warmer, causing more snow and ice to melt. This *ice albedo feedback* is a prime example of a *positive feedback* in the climate system, where a physical process acts to amplify an initial tendency. It is one of the main enablers of climate change. Positive feedbacks act to destabilize or change the climate system, while negative feedbacks act to stabilize or maintain a given climate state.

The albedo of land and vegetation surfaces is ~ 0.15-0.3, but the earth appears to be mostly half clouds and half ocean when seen from space. The albedo of water can be almost 1 when viewed with the sun straight overhead, or if it's setting. But the average albedo of water for all angles of the sun crossing the sky is ~ 0.1. Most of the sunlight that strikes the ocean is absorbed. In fact, fully 1/3 of the sunlight reaching earth goes into evaporating water. Clouds have a typical albedo of ~ 0.5. With a half ocean / half cloud covered earth the average albedo would be 0.3. This is very close to the observed albedo of the real earth as seen from space. This means that 70% of the sunlight reaching our planet is absorbed in the earth system, keeping the earth from eventually cooling off to the universal "background emission temperature" of ~ 4 K. The fraction of energy absorbed is 1-A, so Energy In = S(1-A), and we're almost done.

*5) The earth intercepts sunlight with the equivalent geometry of a circle.* Consider the shape of a shadow cast by a ball illuminated by a flashlight. The amount of light that is "missing" has the area of a circle, not a sphere. This is the amount that strikes the ball, with light rays coming in parallel near the ball's equator, and spreading out across the ball's poles. This explains why the earth's poles are cold and the equator is hot. The area of a circle of radius r is $\pi r^2$, but the area of a sphere of radius r is $4\pi r^2$. Infrared radiation is emitted from the spherical earth in all directions all of the time, so the emitting area is four times larger than the effective absorbing area of the earth for solar radiation. This ratio of 4 does not depend on the size of the object, so the size of an object does not affect its radiative equilibrium temperature. $T_{re}$ is the same for the earth or a tennis ball at the earth's orbit.

Applying these five principles, $S\,(1\,-\,A) = \,4\,\sigma\,T_{re}^4$. Solving for our radiative equilibrium temperature,

$$T_{re} = \left(\frac{S(1-A)}{4\,\sigma}\right)^{1/4} \quad , \qquad (2.7)$$

$T_{re}$ can change if the albedo increases or if the solar output changes. Using $S$ = 1370 W m$^{-2}$ and $A$ = 0.3, $T_{re}$ = 255 K. This is 33 K colder than the observed global and annually averaged surface temperature. This is much colder than 273 K, the freezing point of water. Should the earth be covered in ice? Did we do something wrong? No, the scientific principles and values are correct. We can interpret the result of 255 K as proof of the *atmospheric greenhouse effect*. In the preceding analysis only radiative transfer between the earth and space was considered, not radiative transfer between the earth's surface and the atmosphere. The proper interpretation of the result 255 K is that the average 240 W m$^{-2}$ emission of infrared radiation to space comes from air molecules and clouds at about 5 km altitude, where the temperature is ~ 255 K. The atmosphere also emits infrared energy downward, which is extra energy for the surface, making it warmer than the radiative equilibrium temperature of 255 K.

Application of this equation to Venus and Mars helps to illuminate the greenhouse effect of atmospheric envelopes. Venus is 108 million km from the sun, so from (2.6), $S_{Venus}$ ~ 2600 W m$^{-2}$. Since $A_{Venus}$ = 0.78 due to a thick layer of sulfuric acid droplets, $T_{Venus}$ = 225 K. But the observed surface temperature of Venus is ~ 800 K! Venus' atmosphere is mostly made of $CO_2$ and is 90 times as massive as the earth's atmosphere. The Venutian atmospheric greenhouse effect is ~ 600 K and infrared emission to space occurs from the cold sulfuric acid cloud tops at 100 km altitude at a temperature of 225 K.

Mars is 228 million km from the sun, so from (2.6) $S_{Mars}$ ~ 590 W m$^{-2}$. $A_{Mars}$ = 0.17 due to its reddish soil, so $T_{Mars}$ = 215 K. The observed surface temperature of Mars is ~215 K! Although the Martian atmosphere is mostly made of $CO_2$, it is only 0.7% as massive as the earth's atmosphere. The Martian atmospheric greenhouse effect is minuscule and infrared emission to space occurs primarily from the surface at 215 K. This comparison also suggests the potential for increasing $CO_2$ concentrations to cause an increase in surface temperature on earth.

## 2.8. The Atmospheric Greenhouse Effect

The globally-averaged surface temperature has varied from about 282 to 290 K over the past million years. Although seemingly small, this range of ~8 K has had a large effect on ecosystems and civilization. Even during an ice age, 282 K is still 27 K higher than $T_{re}$ = 255 K. Since there is no way for a planet to gain or lose thermal energy other than by transfer of electromagnetic radiation through space, 255 K is the correct value for $T_{re}$. If you had infrared-seeing goggles on, from space you could detect infrared emitted from a variety of molecules from a range of altitudes centered at about 5 km altitude, where the average temperature is 255 K. This, in fact, verifies the Stefan-Boltzmann law and the other four principles used in deriving $T_{re}$. Does this not lend faith in the method of science?

The earth's atmosphere is mostly opaque in the infrared, but is mostly transparent in the visible. The atmosphere emits infrared energy downward, which keeps the surface and lower atmosphere warmer than from solar radiation alone. This is the natural atmospheric greenhouse effect. Above the tropopause, emission of infrared to space cools the atmosphere, balancing the absorption of UV by ozone. The addition of more gases that emit in the infrared will heat the surface and troposphere and cool the upper atmosphere. The physical principles which allow us to calculate $T_{re}$ apply both to the natural greenhouse effect and to the anthropogenic change in the greenhouse effect.

A fun way to show how the atmosphere transmits solar radiation and absorbs infrared radiation is for one person to play the earth, one to play the sun, and one to stand inbetween and play the atmosphere. The sun throws yellow balls (photons of visible light), which the atmosphere can't catch (transmitting the light), but the earth can catch them (absorb most of the photons) or reflect them (bounce some of them back). The earth can throw red balls (photons of infrared radiation). The atmosphere will let some of them go by out to space (transmit), but will catch (absorb) most of the red balls and throw them back (re-emit) to earth. It will be evident that the earth has more photons to deal with when the atmosphere is added. This causes the atmospheric greenhouse effect: The atmosphere readily *transmits visible* light. The earth *emits infrared* light, but molecules in the atmosphere *absorb* and *re-emit* the infrared radiation back down to the surface. This extra energy raises the temperature of the earth by ~ 34 K, at present. If the atmosphere gets better and better at catching red balls and throwing them back down, then the earth will just have to take the extra heat.

We can distinguish between the radiative nature of the atmospheric greenhouse effect and the way that an actual greenhouse works. A glass-roofed greenhouse lets in visible light, which heats the interior of the greenhouse. The warmed air cannot rise, so the roof acts as a physical barrier to convective heat transport by air motions, and the air inside stays warmer than outside the greenhouse. In the atmosphere, infrared photons are absorbed and re-emitted downward.

Figure 2.2 shows the emission spectra of the earth and the sun, with equal areas to represent equal energy absorbed and emitted in radiative equilibrium. This solar curve therefore corresponds to the solar intensity at 1 astronomical unit, spread over the earth as it rotates, which just balances infrared emission to space. The second curve shows the percent absorption for each wavelength, for photons travelling upward or downward through the entire atmosphere. In the 0.2-0.3 $\mu$m band ozone absorbs most of the incoming solar UV. In contrast, most of the visible light is transmitted through to the surface. Note the absorption lines which begin to appear in the near infrared. The atmosphere absorbs about 23% of incoming solar radiation, largely in the near-infrared part of the spectrum. The atmosphere is even more effective at absorbing infrared photons between 4 and 100 $\mu$m. A minimum in absorption occurs near 8-12

$\mu$m. Since an infrared photon is more likely to reach space in this part of the spectrum it is referred to as the "atmospheric window". Human activities are adding more trace constituents to the atmosphere which are good at absorbing infrared radiation in the atmospheric window, thereby trapping more infrared heat.

Why does the atmosphere have the dual property of readily transmitting visible light, but absorbing most of the infrared? The primary constituents $N_2$, $O_2$, and Ar are tightly bound and have symmetric electron cloud distributions. Visible photons are not energetic enough to break them apart or strip off an electron, so they do not interact with these primary constituents. To absorb or emit infrared radiation a molecule needs to have a bias in electron cloud distribution, or electronic dipole, so that the molecule can change both vibrational and rotational energies simultaneously, as required by the exclusion principles of quantum physics.

$N_2$, $O_2$, and Ar have negligible electronic dipoles, so they are very poor absorbers and emitters of infrared radiation. But molecules such as $H_2O$ and $CO_2$ have dipole moments and many modes of vibration and rotation, making them excellent emitters and absorbers of infrared radiation. Water vapor is a bent triatomic molecule, where the 8 oxygen protons attract the electrons better than the solo hydrogen protons, so there is an electronic dipole, hence energy associated with rotation. $H_2O$ has three primary modes of vibration: bending, symmetric stretch, and asymmetric stretch. Each combination of change in rotational and vibrational energies corresponds to a photon of unique energy, which corresponds to a unique wavelength. Only quantized energy packets can be absorbed or emitted. Infrared absorption spectra are characterized by complex sets of absorption lines that are unique to each molecule.

Since clouds cover about half of the earth and are made of liquid and solid water, they are excellent absorbers and emitters of infrared radiation. They are the primary greenhouse agents in the atmosphere. The current distribution of clouds and cloud types is not known with an accuracy of better than ~10% [e.g., Wylie et al. 2005]. This makes it difficult to quantify the radiative effect of clouds at present. It is even more difficult to be certain about the distribution and type of clouds in the future. The feedbacks of clouds in a greenhouse warming scenario represent a significant uncertainty in forecasting future climate.

The second most important greenhouse constituent is water vapor. In warm climates, it can compose as much as 4% of the air. Although an individual water molecule is only moderately efficient at absorbing and emitting IR, there are so many water molecules that, in clear air, the distribution of water vapor largely determines the greenhouse effect. The stratosphere is very dry, so if an infrared photon is emitted above the moist troposphere, it is much more likely to reach space. This is shown in the bottom curve in Fig. 2.2. In the stratosphere, $CO_2$ dominates absorption and emission of infrared. There an approximate balance between ozone absorption of UV and $CO_2$ emission of infrared holds. As more $CO_2$ is added to the atmosphere, the colder the stratosphere becomes, due to greater infrared emission to space.

$CO_2$ is a linear molecule, but can generate robust electronic dipoles when in its bending mode or asymmetric stretch mode. It is sufficiently complex that it has many possible discrete energy changes available for absorption and emission line spectra. The 15 $\mu$m absorption band for $CO_2$ is shown in Fig. 2.9, which results from changes in vibrational energy states enabled by temporary electric dipoles in the bending mode. Spectral lines occur in three groups, which correspond to families with changes of -1, 0, or +1 in the rotational quantum state, with an individual line representing a quantized change in energy state. By adding more $CO_2$, absorption (and emission) will increase in the 15 $\mu$m band, leading to an enhanced greenhouse effect.

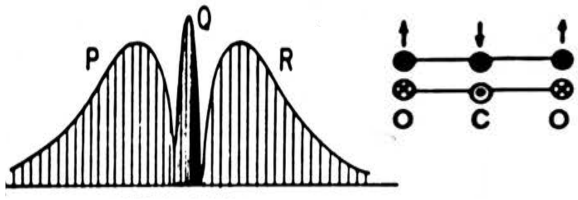


**Figure 2.9.** Vibration-rotation absorption bands in the 15 $\mu$m infrared band of carbon dioxide [Andrews et al. 1987].

After clouds, water vapor, and $CO_2$ comes a host of complex molecules generated by natural and human processes, including $N_2O$, CFCs, and $CH_4$. The variety and concentration of new manufactured chemical compounds which are good greenhouse gas molecules is growing rapidly, with hundreds of new ones each year, proliferating into the thousands. This collection represents the second tier of anthropogenic greenhouse agents, after increases in $CO_2$.

It is also of interest to point out that $O_3$ is not only a good absorber of UV, it absorbs and emits infrared nicely near 9.6 $\mu$m, making it a significant greenhouse gas. The loss of 6% of our ozone globally over the past few decades has contributed to stratospheric cooling, but has also reduced downward emission at 9.6 $\mu$m to the surface. Hence stratospheric ozone depletion is mildly offsetting anthropogenic greenhouse warming.

*Focus: Atmospheric Emissivity*

The equation for $T_{re}$ can be modified to include the extra emission of infrared by the atmosphere down to the surface: $\epsilon\,\sigma\,T_a^4$, where $T_a$ is the atmospheric temperature, and $\epsilon$ is the *atmospheric emissivity*, by adding this on the right-hand side of (2.5):

$$\sigma\,T_s^4 = \frac{(1-A)\,S}{4} + \epsilon\,\sigma\,T_a^4\,, \qquad (2.8)$$

where the earth's surface temperature, $T_s$, is now distinguished from the radiative equilibrium temperature. With $A = 0.3$, $S = 1370$ W/m$^2$, and $T_a = 255$ K, and a heuristically chosen value of $\epsilon = 0.63$, $T_s = 288$ K, which is close to the observed temperature of the earth. In this simple model, if a continued rise in anthropogenic greenhouse gases led to a 20% increase in atmospheric absorptivity, so that $\epsilon = 0.77$, $T_s = 292$ K, an increase of 4 K. This is consistent with forecasts for the end of this century using state of the art climate models.

Human activities are increasing the quantities of gases which absorb actively in the 8-12 $\mu$m band, "smudging" the atmospheric window, so that more infrared photons are absorbed, making the surface warmer. Greenhouse gases act as positive feedbacks or amplifiers in the climate system. During cold times, $CO_2$, $CH_4$, and other gases tend to stick to and dissolve into the ocean and freeze into high latitude soils, thereby decreasing the greenhouse effect, and allowing the planet to get colder. During warm times these constituents tend to come out of the ocean and soil into the gas phase, amplifying the greenhouse effect and furthering the warming. $H_2O$ tends to stay in the ocean and in ice during cold times, reducing the greenhouse effect, and tends to evaporate more into the air during warm times, amplifying the greenhouse effect. The positive feedbacks due to greenhouse gases and the effect of ice albedo are essential ingredients for fostering transitions between glacial and interglacial states.

In a global warming scenario, how do changes in clouds affect global temperatures? Increased surface temperatures lead to enhanced evaporation, which will lead to more clouds. If deep convective clouds increased, there would be more high cirrus detrained from thunderstorm anvils. Thin high cirrus clouds trap infrared radiation and let in enough sunlight so that there is a net warming effect. Increased high clouds would have a positive feedback on global warming. If enhanced evaporation leads mostly to more low clouds, which are warm and thick, they would trap very little infrared but reflect more sunlight, leading to a cooling, or a negative feedback. This uncertainty regarding the response of clouds to climate change and their feedback on the climate system is the basis for vigorous scientific discussion. Most numerical climate simulations suggest a moderate positive feedback from clouds in the current climate state.

The most robust anticipated changes are that warmer sea surface temperatures (SSTs) will cause more $H_2O$, $CO_2$, and $CH_4$ to go into the air from the ocean and soils. At higher latitudes soil microbes will oxidize ancient organic matter, emitting more $CO_2$ and $CH_4$. Ice masses will melt. The ice albedo feedback and greenhouse gas feedback will amplify the warming. There is some concern that these positive feedbacks may lead us into a new climate state, perhaps a "*super-interglacial*", or something approaching the balmy Cretaceous, some 100 Mybp. It is possible that there is an upper limit due to more clouds reflecting enough sunlight.

Some of the primary concerns about the anthropogenic greenhouse effect include changes in the large-scale weather patterns and in the frequency and intensity of storms, increased frequency of heat waves, changes in the ocean circulation, rapid polar warming, rise in sea level, synergy with disease organisms, changes in rainfall patterns challenging crop growing strategies, and the potential to accelerate species loss due to inability to adapt. Since 1880 sea level has risen ~ 25 cm due to thermal expansion of the surface waters and melting of ice and snow on land. It is expected to rise at least 50 cm more in this century, inexorably increasing pressure on the enormous population living near sea level. One possible savior could be the effect of warming the circumpolar Antarctic on increasing glacial mass. Since the surface is at such a high altitude and so cold, a minor rise in temperature may provide more moisture, which would increase snowfall. Another interesting aspect of Antarctica is that the current springtime ozone hole implies diminished downwelling 9.6 $\mu$m infrared radiation due to the diminished ozone. *Thompson and Solomon* [2002] argue that Antarctic warming is being delayed by the presence of the ozone hole. It is anticipated that global warming will begin to influence Antarctica more as the ozone hole fills and emits more downwelling infrared radiation to its surface. The future trajectory of the Antarctic ice budget is uncertain.

Figure 2.10 shows the global average radiative forcing in W $m^{-2}$ at the surface from different anthropogenic changes in atmospheric composition since 1750. $CO_2$ is the largest single factor at 1.7 W $m^{-2}$. Downwelling radiation has increased from the second group including $CH_4$, $N_2O$, and halocarbons (chlorine and bromine compounds) by another 1.0 W $m^{-2}$. The reduction of stratospheric ozone has reduced downwelling infrared by 0.05 W $m^{-2}$, but increases in tropospheric ozone have increased it by 0.35 W $m^{-2}$. The stratosphere is getting moister, as rising $CH_4$ is photolyzed, with some of it being converted to $H_2O$, with the resulting minor increase in downwelling infrared approximately offsetting the reduction due to stratospheric ozone decrease. Changes in land use have increased the surface albedo, but this is partly offset by the reduced surface albedo of snow due to soot. Anthropogenic aerosol, including sulfuric acid droplets produced from burning fossil fuels, reflect sunlight, decreasing downwelling solar radiation by 0.5 W m$^{-2}$. Although there are significant uncertainties, anthropogenic aerosols tend to increase the number of nucleated droplets and ice crystals in clouds such that they last longer and reflect more sunlight, reducing downwelling solar radiation by 0.7 W m$^{-2}$. Jet contrails tend to cool the daytime by reflecting solar radiation and warm the nighttime by trapping infrared radiation, but the net effect is small. The total net anthropogenic forcing for all of these factors is 1.6 W m$^{-2}$.

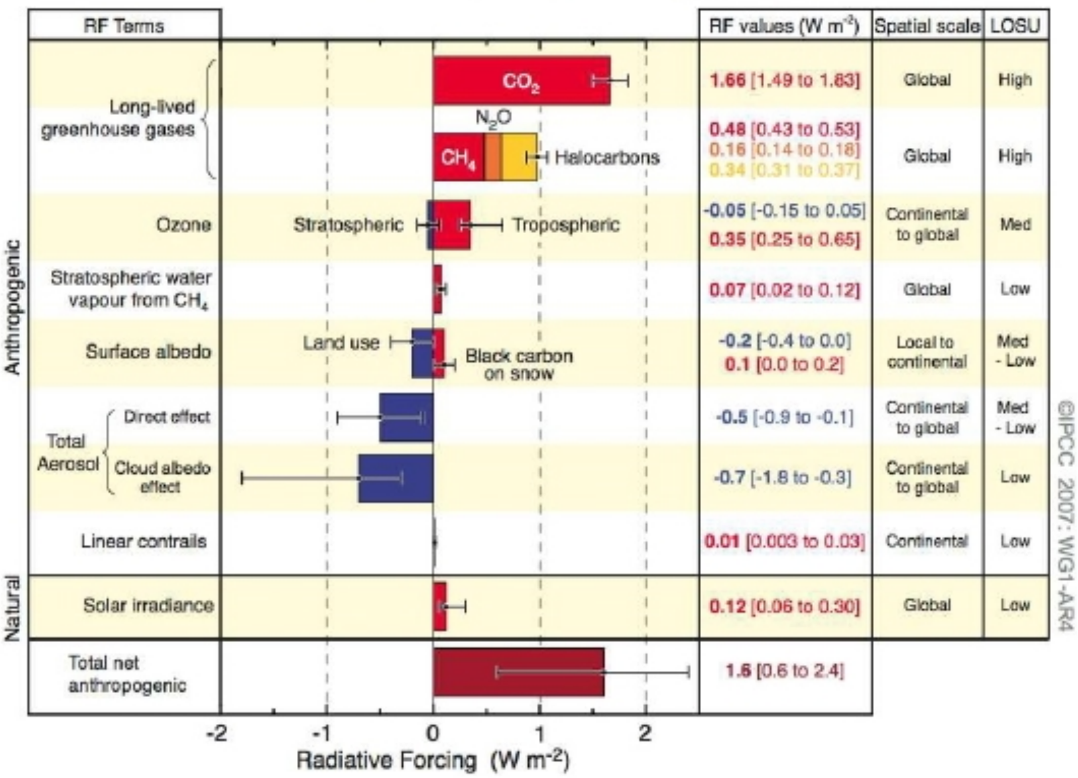


**Figure 2.10.** Global average radiative forcing estimates (color bars) and uncertainties (solid bars) in 2005 for anthropogenic $CO_2$, $CH_4$, $N_2O$ and other important agents and mechanisms, excluding volcanic effects. [Fig. SPM-2, IPCC 2007].

It is of interest to compare this with changes in the output of the sun. The IPCC report shows that since 1750 the output of the sun has increased very slightly, implying an increase of 0.1 W m$^{-2}$ at the surface. But this analysis does not take into account the regional effects of aerosols. Many studies have reported reductions in the solar intensity at surface stations around the world in and downstream of populated regions. From 1900 to 1980 cloud-free solar intensities diminished by ~ 10-40 W m$^{-2}$, the so-called solar dimming. Since 1980 this trend has reversed itself. Increasing anthropogenic aerosol, a byproduct of burning fossil fuels, absorbed increasing amounts of sunlight until around 1980, when catalytic converters, smokestack scrubbers, and other technologies helped to alleviate aerosol pollution in many industrialized regions. These values are much larger than the 1.6 W-m$^{-2}$ increase noted above. The uncertainty in future aerosols is as large as that associated with the cloud response to increasing water vapor.

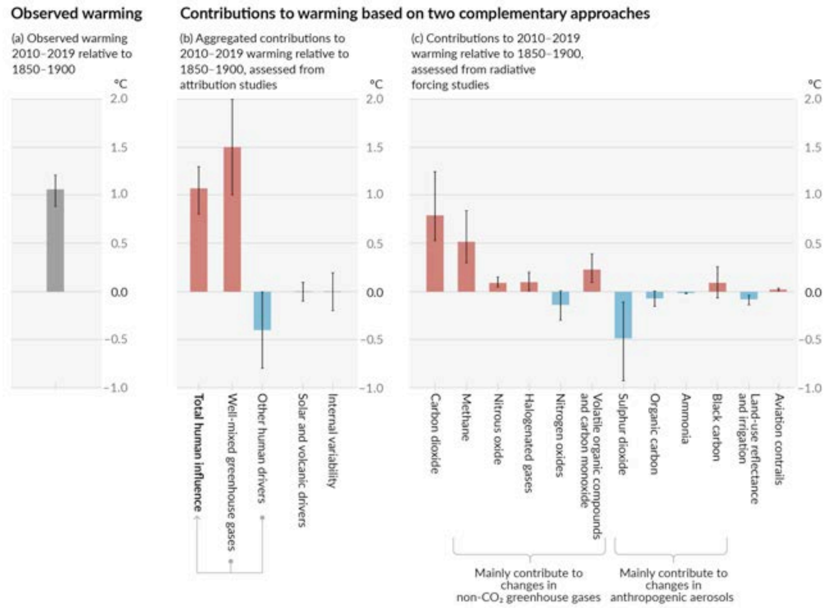


**Figure 2.11.** Contributions to observed warming in 2010–2019 relative to 1850–1900. a) Observed increase in global surface temperature. b) Temperature change attributed to: total human influence; greenhouse gases; aerosols, ozone and land-use change; solar and volcanic drivers; and internal climate variability. c) Temperature changes from individual components of human influence. Whiskers show very likely ranges. For aerosols, both direct effects (through radiation) and indirect effects (through interactions with clouds) are considered. [IPCC Sixth Assessment Report, 2022, Figure SPM.2].

Figure 2.11 shows the contributions to temperature change from 1850 to the present, from the IPCC Sixth Assessment Report (2022). Their primary conclusion is that observed warming is driven by human activities, with greenhouse gas warming partly masked by aerosol cooling.

In summary, the earth system is warming in conjunction with understood and quantifiable processes. Specific uncertainties include the response of Antarctica and clouds, the effects of aerosols, changes in solar output, decadal volcanic clustering, and the potential for nonlinear, unanticipated responses in the complex climate system. A consistent picture of anthropogenic global warming is beginning to be quite compelling. To understand the pattern of global warming we need to better understand the properties of the earth system, especially global patterns of circulation in the atmosphere and ocean. These are also useful for understanding the distribution of various pollutants.

**Key Terms**

*absorption* – when electromagnetic radiation is absorbed by a substance (energy increase)
*atmospheric greenhouse effect* – absorption of infrared by clouds and constituents in the atmosphere and re-emission downward increases lower tropospheric temperatures
*coccolithophores* – microscopic photosynthetic organisms at the base of the food chain in the ocean, which create their shells from dissolved calcium carbonate, making them susceptible to acidification
*Coriolis effect* – our reference frame on the earth's surface accelerates around the rotation axis, so straight-line motion appears to be curved, to the right (left) in the NH (SH)

*diatoms* – microscopic photosynthetic organisms at the base of the food chain in the ocean which create their opal shells from dissolved $SiO_2$
*electromagnetic spectrum* – the range of wavelengths of light from extreme uv to radio waves
*emission* – when electromagnetic radiation is emitted by a substance (energy decrease)
*emissivity* – the ratio of emitted electromagnetic radiation to that which would be emitted by a perfectly emitting substance or "black body"
*exponential profile* – a mathematical expression for describing a quantity whose rate of change increases with increasing distance
*geostrophic balance* – for large-scale flows, acceleration is weak compared to the pressure gradient force, which is balanced by the Coriolis effect
*hydrostatic balance* – when the force of gravity is balanced by the upward pressure gradient force, a very good approximation for most of our fluid envelopes
*hydroxyl radical* – a highly reactive, short-lived compound which regulates partitioning among chemical species and helps clean air of pollution
*ice albedo feedback* – an important positive feedback in the climate system, where more ice reflects sunlight, which cools, allowing for more ice, and less ice absorbs sunlight, which warms, leading to less ice
*longwave radiation* – infrared radiation, longer than ~4 $\mu$m, characterizing electromagnetic radiation emitted at typical temperatures in the earth system
*negative feedback* – when the climate system responds to a perturbation by returning to its original state
*positive feedback* – when the climate system responds to a perturbation by amplifying it, whether the perturbation is positive or negative
*radiative equilibrium* – when the amount of electromagnetic radiation being absorbed by a body equals the amount being emitted
*radiative equilibrium temperature* – the temperature that a body will have in radiative equilibrium
*radiosonde* – a meteorological instrument package attached to a balloon which measures the atmosphere above the surface, with resulting data guiding weather forecast models
*ridge* – an atmospheric pattern where the midlatitude westerlies are diverted poleward around a region of high pressure
*shortwave radiation* – visible radiation, shorter than ~4 $\mu$m, characterizing electromagnetic radiation emitted at typical temperatures in the photosphere of the sun
*solar constant* – average strength of solar radiation at earth's orbit, ~ 1370 W $m^{-2}$
*superinterglacial* – hypothesized future climate state resulting from anthropogenic global warming which notably exceeds interglacial temperatures in the recent million years
*synoptic charts* – large-scale maps compiling weather observations taken all at the same time
*teleconnections* – geographical patterns of co-variation across a wide range of the globe
*transmission* – the amount of electromagnetic radiation being transmitted by a substance, instead of being reflected or absorbed
*tropospheric cleanser* -- the hydroxyl radical helps pollutants rain out
*trough*– an atmospheric pattern where the midlatitude westerlies are diverted equatorward around a region of low pressure

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## Quantitative Problems

1. Given that the density of the human body is similar to that of sea water, what radius is required for a balloon to contain all humans on the planet?

2. a) During the first billion years of the earth's history, the solar output was only about 75% of its modern value. Estimate the radiative equilibrium temperature of the earth, T, keeping A=0.3, for this ancient situation.

b) A large volcanic eruption or large continental glacier could increase the earth's albedo. Estimate T for A=0.4, keeping S=1370 W/m$^2$.

3. If the earth's orbit were at a different distance from the sun (r′), instead of the present value (r), we can estimate the solar intensity at the new distance (S′) from $S'/S = (r/r')^2$, where S=1370 W/m$^2$ and r = 150 x 10$^6$ km. How many kilometers *farther* from the sun (r′ - r) would the earth's orbit have to be for T to be 15 K colder than at present?

4. If the average surface temperature of the earth is 288 K and the radiative equilibrium temperature of the earth is 255 K, estimate the intensity of radiation emitted from the earth's surface and from the earth to space in W/m$^2$.

5. If all of the radiation absorbed by the planet went into evaporating water, how much water could be evaporated in one day from a typical square meter of ocean surface. [1 day ~ 10$^5$ s; it takes 2.5 x 10$^6$ J to evaporate 1 kg of water; water density is 1000 kg/m$^3$].

6. Using Wien's Law estimate the wavelength of maximum emission from the surfaces of a) Mars, b) Earth, c) Venus, d) your skin. You will need to use Kelvins and microns.