1 EARTH'S CLIMATE SYSTEM

Earth's climate system includes all the realms of the planet that interact to produce the seasonal march of temperature, wind, and precipitation. Most important are the atmosphere; the oceans, including their linked chemical and biological processes; and the solid Earth insofar as it influences CO_2 concentration in air. Atmospheric processes govern climate over time scales of a few years or less. The oceans influence climate change over periods of decades to tens of millennia. Over periods of a hundred thousand years or more, interactions between the solid Earth and the surface environment fix the CO_2 concentration of air and Earth's average temperature.

In this chapter, we discuss the physical and chemical controls that determine the most important characteristics of Earth's climate. We start by discussing the decrease of pressure and temperature with elevation. We proceed to Earth's heat budget and the controls on global average temperature. We then examine the large scale circulation of Earth's atmosphere, and discuss how this circulation dictates prevailing wind directions at the surface and how it determines what regions of the globe get a lot of precipitation and what regions are dry. We discuss ocean circulation, biological processes in the ocean, and how these processes combine to change the partial pressure of CO_2 in the atmosphere over periods of decades to millennia. We end with a brief description of the geological processes that fix the average background CO_2 concentration of the atmosphere over hundreds of thousands of years.

ATMOSPHERIC PROPERTIES AND CLIMATE

Pressure and temperature as a function of altitude

Earth is heated by sunlight predominantly at ground level, which in turn warms the local lower atmosphere. Warm air expands and becomes buoyant, leading to vertical mixing. As air rises, it encounters lower pressures and expands into the void. Atmospheric pressure decreases with elevation in a way that reflects hydrostatic equilibrium in the atmosphere. In this condition, air is stabilized at a given altitude by the balance between gravity, which pulls the air mass down toward the surface, and the upward push exerted by the natural tendency of a gas to expand.

At a given elevation, pressure is simply the weight per unit area of the overlying column of air. This condition is expressed by the equation:

$$d\rho/dz = -g\rho = M_{air}/RT, \tag{1}$$

where ρ is density, z is height above the surface, g is gravitational acceleration, M_{air} is the molar mass of air

(29 gm/mole), R is the ideal gas constant, and T is Kelvin temperature. From this equation, one can show that pressure decreases by a factor of 1/e (0.37) for every ~7–8 km increase in elevation for typical air temperatures.

Temperatures are cold at higher altitudes because of the decrease of pressure with elevation. Consider a parcel of dry air large enough that it is not gaining or losing heat to the surrounding atmosphere. As this parcel rises, it encounters lower atmospheric pressure and "pushes out" into the surrounding air. In so doing, it uses energy, which leads it to cool. The cooling rate, or "lapse rate," is about 10°/km. If the air is wet, water condenses as it rises and reaches the dew point. Latent heat is released, and the lapse rate is smaller, typically 4–7°/km. Lower values correspond to warmer saturated air, with more water vapor and greater potential to release latent heat.

This decrease with temperature reverses at an altitude of about 11 km. The reversal is caused by the absorption of high energy (ultraviolet or UV) light from the sun due to reactions of the ozone cycle. These reactions are:

$$O_2 \rightarrow 2\,O \tag{2}$$

$$O_2 + O \rightarrow O_3 \tag{3}$$

$$O_3 \rightarrow O_2 + O \tag{4}$$

$$O + O_3 \rightarrow 2\,O_2 \tag{5}$$

Absorption of ultraviolet light by O_2 (oxygen) and O_3 (ozone) has the net effect of warming the surrounding air. It is this warming that causes the temperature to increase with altitude above about 11 km elevation. This increase in temperature continues to an elevation of ~50 km, where pressure is about 1% of the sea level value. Above 50 km, the temperature again begins to fall because O_2 is not abundant enough to allow significant O_3 production. There is an additional reversal at about 85 km elevation.

The *troposphere* is the atmospheric layer from the surface to the first temperature minimum, and the surface of minimum temperature is the *tropopause*. The *stratosphere* is the overlying layer of air from about 11 to 50 km elevation. These are the two lowest layers of the atmosphere.

Solar heating and radiative equilibrium

At the top of the atmosphere, the cross sectional area of Earth receives heat from the sun at the rate of 1368 watts m^{-2}. Spread over Earth's entire daytime and nighttime spherical surface, the average heating is 4 times lower, at 342 W m^{-2}. Some of this heat is reflected back to space. The remainder is redistributed in various ways between the land surface, ocean, and atmosphere. However, it is lost only by radiation of photons or electromagnetic radiation to space. The loss is described by the Stefan-Boltzmann equation:

$$\text{Rate of energy loss} = \sigma T^4, \tag{6}$$

where T is Kelvin temperature and σ is the Stefan-Boltzmann constant, 5.67×10^{-8} w m^{-2} K^{-4}.

Earth's average surface temperature changes only very slowly with time, so that heat must be lost at nearly the same rate at which it is received. Equation (6) can then be rearranged to solve for temperature. The temperature calculated for a heat flux of 342 W m^{-2} is 6°C, not too different from Earth's preindustrial average surface temperature of about 15°C. Unfortunately, there are two serious omissions in this calculation. The first becomes apparent by examining figure 1.1: much of the light reaching the Earth is simply reflected back to space, without ever warming the surface and contributing to Earth's heat budget. Sunlight is reflected by all surfaces, but the brightest (most reflective) are clouds, snow and ice, and deserts. Earth's global reflectance, or *albedo*, is 0.31. Correcting for albedo, a value of −19°C is calculated, which is ~36°C too low for Earth's surface temperature. What's wrong?

Actually, nothing. A value of −19°C is Earth's radiative equilibrium temperature, but it is not achieved at the surface. The reason for this is the greenhouse effect, which is illustrated in figure 1.2. The top panel (a) shows the energy density of solar and Earth radiation as a function of wavelength. Wavelength increases to the right; frequency, and energy of electromagnetic radiation, increase to the left. Because the surface of the sun is so hot (~6000 K), most solar energy is radiated in the visible region of the electromagnetic spectrum. Radiation from the cool Earth, on the other hand, is in the longer wavelength, lower energy infrared region. Panels (b) and (c) show the fate of radiation as it passes through the

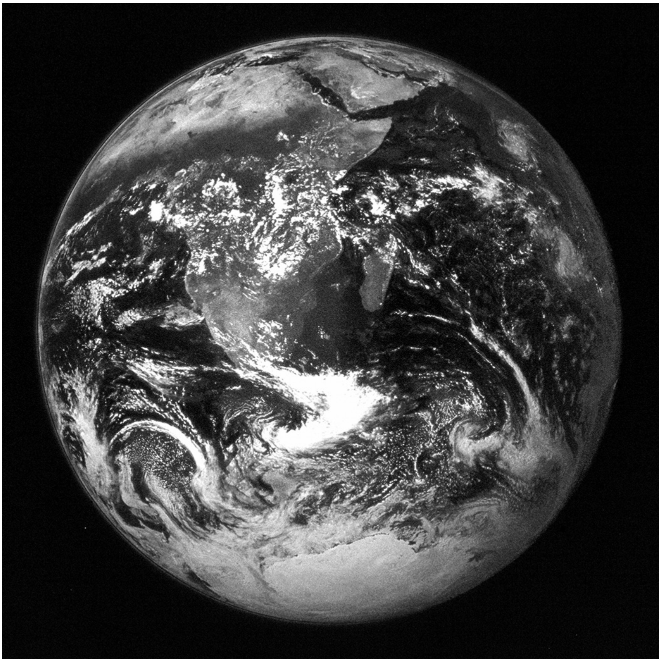
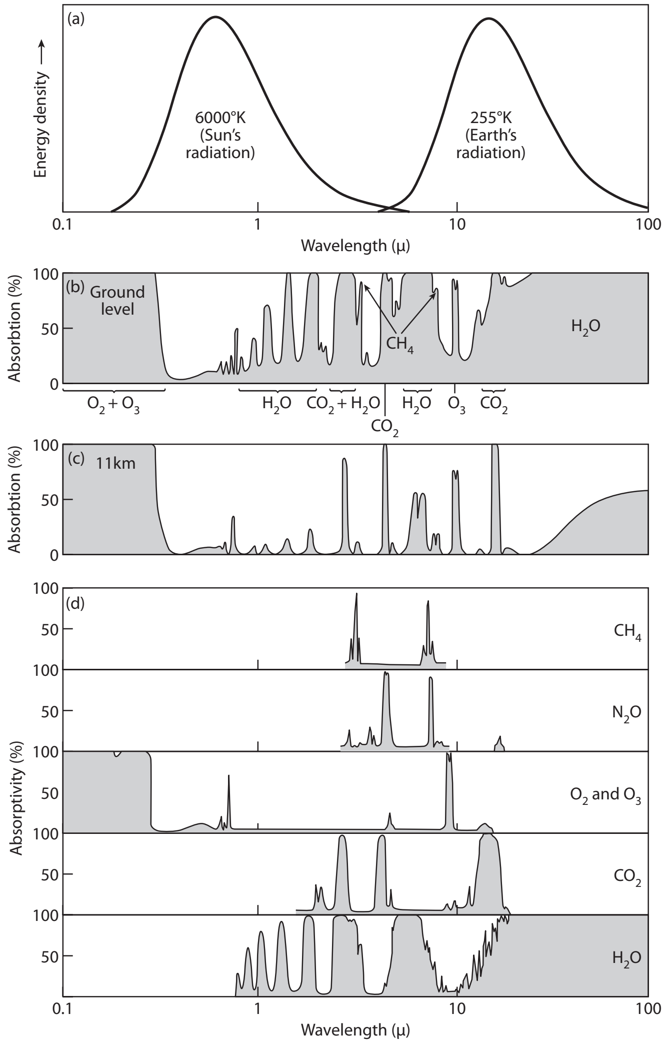


Fig. 1.1. Earth viewed from Apollo 17 in space. Landmasses from bottom: Antarctica (ice-covered), and Africa, with Madagascar to the east, and the Saudi Peninsula to the northeast. The Southern Ocean separates Antarctica from South Africa, the Atlantic is to the west of Africa, and the Indian to the east. Ice-covered Antarctica and clouds are the most reflective surfaces; the North African desert and Saudi Peninsula are next. The forests of tropical Africa are even less reflective, and the ocean is the least reflective realm. This image illustrates that much incoming sunlight (31% globally) is reflected back to space.

atmosphere. In these panels, white means that radiation is transmitted, and gray indicates that it is absorbed by interactions with molecules of the gases in air. Absorbed radiation is used to kick electrons into higher energy levels, and to increase vibrational and rotational frequencies of molecules. Panel (b) shows the absorption of radiation as a function of wavelength between the surface and the top of the atmosphere. Most solar radiation passes through the atmosphere "intact"; it is this property that allows us to view the sun, Moon, and stars. Most Earth radiation, on the other hand, is absorbed as it passes through the atmosphere. Absorption warms the air and leads to reradiation of the absorbed energy. Some of this reradiated energy is transmitted downward toward the surface, where it delivers an extra serving of heat. Thus, the surface is warmer than it would be if absorbing (or greenhouse) gases were absent from the atmosphere.

Panel (c) shows the fraction of radiation absorbed between 11 km and the top of the atmosphere. In this interval, almost all outgoing Earth radiation is transmitted, and there is no longer much warming of the atmosphere due to absorption of outgoing infrared radiation.

Ideally, there would be some level in the atmosphere below which infrared radiation is largely absorbed, and above which it is mostly transmitted. It is at this hypothetical level that Earth attains its radiative equilibrium temperature of −19°C. This level is at about 5 km elevation. The average lapse rate in the troposphere is about 6.5°C/km, so that temperature rises by 33° from the


(a)
Energy density
6000°K
(Sun's
radiation)
255°K
(Earth's
radiation)
0.1
1
10
100
Wavelength (μ)
(b) Ground
level
Absorbtion (%)
100
50
0
CH_4
H_2O
$O_2 + O_3$
H_2O
$CO_2 + H_2O$
CO_2
H_2O
O_3
CO_2
(c) 11km
Absorbtion (%)
(d)
Absorptivity (%)
CH_4
N_2O
O_2 and O_3
CO_2
H_2O
Wavelength (μ)

radiative equilibrium level to sea level. The calculated average sea level temperature is 14°C, close to the observed global average. The concept of a single level where the infrared radiation ceases is greatly simplified, but the idea is correct.

Panel (d) illustrates the absorption of radiation by different gases between the surface and the top of the atmosphere. In the ultraviolet, all absorption is due to O_2 and O_3, illuminating the role of ozone as the UV shield. In the infrared (IR), most absorption of radiation is due to water. Of the other so-called greenhouse gases, CO_2 is by far the most important absorber, followed by CH_4 (methane) and N_2O (nitrous oxide). Ozone also absorbs in the IR, and there is thus a small contribution to the greenhouse effect both from tropospheric and stratospheric ozone.

Fig. 1.2. (*a*) Energy density of radiation given off by black bodies at 6000 K (representing the sun) and 255 K (representing Earth) as a function of wavelength in micrometers. (*b*) Percentage of radiation absorbed while passing between the top of the atmosphere and the surface of the Earth. Most solar energy is transmitted, while most outgoing Earth radiation is absorbed. (*c*) Percentage of radiation absorbed between the top of the atmosphere and 11 km elevation. Most solar and Earth radiation is transmitted by the thin (and dry) atmosphere in this region. (*d*) Contributions of the different gases to absorption of radiation between the top of the atmosphere and the surface of the planet. Water is the most important greenhouse gas, followed by CO_2. From Peixoto and Oort 1992.

CHAPTER 1

ATMOSPHERIC CIRCULATION

Sunlight warms Earth's surface, which in turn warms the atmosphere. Under these conditions, one might expect a meridional (latitudinal) circulation system, with warm air rising at the equator, and cool air sinking at the poles. There would be poleward flow aloft, and equatorward flow at the surface. In fact there are convection cells in the atmosphere, but they are not quite this grand. In the "Hadley cell," air rising at the equator flows to a latitude of about 30°, sinks to ground level, and flows back toward the equator. In the "Polar cell," air rises at about 60° latitude, flows toward higher latitudes, sinks at the poles, and again closes the loop by equatorward flow at the surface.

Because of an effect known as the Coriolis force, winds are westerly (from the west) in the upper, poleward flowing air of the Hadley cell. The air flowing poleward must, in the absence of friction, conserve its angular momentum (mass × velocity × radius). Air at the equator is moving toward the east at about the same rate that the surface is spinning on its axis. As this equatorial air rises and moves poleward, its eastward velocity increases in an absolute reference frame because the radius of the Earth is decreasing. With respect to the underlying ground, its velocity is increasing even faster, because the eastward velocity of the ground becomes smaller as the air moves to higher latitudes. From our perspective, with a reference frame rotating along with the Earth, an air mass traveling toward the pole will accelerate toward the east. The Coriolis force is the virtual force producing this acceleration.

Think about air at the equator in relation to the ground. In an absolute frame of reference, it is moving to the east at a velocity of 1671 km/hr, due to rotation of the solid Earth. In the absence of friction and turbulence, this air would be flowing to the east 484 km/hr faster than the ground when it reaches a latitude of 30°. In practice, there is such a feature in the upper troposphere: the jet stream, which flows eastward but at a lower velocity of 150–200 km/hr. Wind speeds must be low near the surface due to friction, and even above the surface, friction and other influences make wind speeds lower than those calculated when only considering conservation of angular momentum.

High velocities aloft, coupled with the decrease in temperatures between the Hadley cell and the Polar cell (at latitudes of roughly 30–60°), lead to a very different atmospheric circulation in these middle latitudes. Here, the circulation is much more chaotic, dominated by cyclones—large air masses rotating counterclockwise in the Northern Hemisphere and clockwise in the Southern Hemisphere. Cyclones travel to the east, leading to transitional regions known as fronts and the variable weather so characteristic of the midlatitudes. Along with the movement of cyclones, there are flows of warm air masses toward the poles and cool air masses toward the equator. These flows lead to the transport of heat from the tropics toward the poles, and attenuate the meridional temperature gradient. For dynamic reasons, there tends to be net sinking of air at the southern boundary of the midlatitude zone (around 30°) and rising air at the northern end (60°).

With this background it becomes fairly straightforward to understand the distribution of surface winds. In the region of the Hadley cell (0–30°), air rises at the equator. Aloft, it flows toward the pole and to the east. At ground level, there is a return flow. Winds turn toward the right (west) as they pass from a region where the surface is turning more slowly to a region where it is turning faster (the equator). The tropics are therefore regions of easterly winds (from the east) known as the trade winds. Between 30 and 60° latitude, there is a net westerly flow, upon which are superimposed the airflows associated with rotating cyclones. Hence, surface winds in this region are variable, despite the mean flow to the east. North of 60°, circulation of the polar cell is similar to that of the Hadley cell: air rises at the low-latitude boundary, flows toward the east aloft, sinks at the poles, and the surface return flow is again easterly.

This dynamic picture also explains the distribution of precipitation. Air rises at the equator and at 60° (on average), then sinks at around 30° and at the poles. When air rises, there are two changes that influence the degree of saturation of water vapor, and hence the amount of precipitation. First, rising air expands, decreasing the concentration of water vapor relative to the saturation concentration at which liquid will form. Second, rising air cools, lowering the equilibrium water vapor concentration. It turns out that the effects of cooling exceed those of expansion, and the degree of saturation of water vapor increases as air rises. Consequently, there is a belt of heavy precipitation along the equator. Sinking leads to the opposite effect: air warms, and its ability to hold moisture rises. This effect leads to the areas of low precipitation in the Northern Hemisphere around 30° N, accounting for the deserts of North Africa and the western United States, and dry climates elsewhere in the subtropics. In midlatitudes, fronts lead to rising air masses and abundant precipitation. Precipitation is low at very high latitudes because cold air can hold very little moisture.

Superimposed on these global patterns are important local features. Perhaps the most important are temperature and precipitation gradients over land produced by interactions between land and the nearby sea. Heat received by land is absorbed at the very surface and transmitted to depth by conduction, which operates very slowly and induces seasonal temperature cycles to depths of only a few meters. Heat received at the sea surface penetrates more deeply and is mixed rapidly to depths of tens to hundreds of meters. In other words, the land surface shares its heat to only a meter or so depth, while the ocean surface shares its heat to 100 m or more. Consequently, seasonal heating and cooling of land is much greater than that of the oceans. This feature manifests itself dramatically in two ways. First, in temperate and subpolar continental areas, seasonality is much stronger in the east than in the west. For example, Baltimore is only 2° further north than San Francisco, but its annual temperature cycle is four times larger (25°C range of monthly average temperatures compared with 6°). These zonal gradients are a sign of the prevailing westerly airflow in the middle latitudes. Air in San Francisco originates from the Pacific Ocean, and reflects its attenuated seasonal temperature cycle. On the other hand, air in Baltimore has crossed the continent and has acquired the large seasonal temperature fluctuations in the center of North America.

Another important local feature is the intense summertime heating of air over land. This makes air buoyant, causes it to rise, and draws in more wet air originating over the oceans. As air rises, it of course cools. Water condenses and falls as rain. This phenomenon gives rise to the monsoons, which involve intense summertime precipitation over large continental areas with meteorological links to the oceans.

THE OCEANS

Ocean circulation

The same laws of fluid flow govern the circulation of the oceans and the atmosphere. However, ocean circulation is very different from atmospheric circulation for two reasons. First, the oceans are heated from the top rather than from the bottom. Second, seawater is more dense than air; thus, ocean currents are much slower than winds. Like the atmosphere, the ocean is dynamic, although it mixes on far longer timescales (about one millennium vs. one year). Three factors cause the oceans to move or mix. First, waters that are more dense than their surroundings will sink, and more buoyant waters will rise. Second, winds transfer momentum to the sea surface, inducing lateral flow and in some cases vertical motions. Third, ocean tides and currents induce vertical mixing of waters in the ocean interior, especially over mountainous areas of the seafloor.

The deeper waters of the oceans originate as the densest waters at the ocean surface. The density of surface waters depends on both temperature and salinity. Surface waters of the oceans are heated in the low latitudes. In the high latitudes, they are cooled by interactions with the atmosphere during much of the year. Therefore, higher latitude waters are colder, as most of us have experienced, which increases their density. Surface waters are also subject to evaporation and precipitation. Evaporation removes water without removing salt, increasing salinity—and density. Precipitation has the opposite effect. Precipitation exceeds evaporation along the equator, and evaporation is more rapid in the subtropics (as on land, where the tropics are filled with rainforest, and the subtropics host the world's great deserts). Consequently, the salinity of seawater tends to be low on the equator, high in the subtropics (about 30° latitude), and lower toward the poles. Since salty water is denser than fresh, salinity variations raise the density of water in the subtropics and decrease it at low and high latitudes. Over the ambient range of surface ocean conditions, density changes due to temperature are roughly twice those due to salinity. Hence temperature wins the competition to influence density; the densest waters are found in the high latitudes, and it is here that waters sink to fill the deep oceans. Even in high latitudes, however, salinity has some impact. North Atlantic salinities are the highest of the polar regions, Southern Ocean salinities are lower, and North Pacific salinities are the lowest. It turns out that North Pacific densities are too low to allow deep water formation in this region. Consequently, it is the North Atlantic and Southern Ocean where waters form that fill the ocean basins below about 1000 m depth. For reference, the average deep ocean depth is ~3800 m.

Winds impart momentum to the sea surface, leading to the flow of water. The consequences are, however, somewhat surprising, because Earth's rotation leads flows to bend to the right in the Northern Hemisphere and to the left in the Southern Hemisphere. Winds blow from the east in the tropics and from the west at midlatitudes. These winds cause waters to flow to the north in the tropics and to the south in the midlatitudes. Waters thus "dome" in the center of the ocean basins, exerting a pressure gradient leading to a circular, or "gyre" flow. These waters circulate in a counterclockwise direction in the Northern Hemisphere and clockwise in the Southern Hemisphere.

Winds also join with density flow and interior mixing (or turbulence) to drive the exchange of waters between the surface and the abyss. Westerly winds blowing over the Southern Ocean drive the eastward-flowing Antarctic Circumpolar Current, which circles the continent. These also drive a flow to the left (north) in roughly the northern half of the Southern Ocean. This northward flow leads to a water deficit in the center of the basin, which induces the upwelling of waters from the deep ocean. Upwelling in turn activates the formation of new deep waters in the North Atlantic and the Southern Ocean. Also contributing to deep water formation is turbulent vertical mixing driven by tides and currents flowing over rough bottom topography. In this process, heat is mixed from shallow depths into the deep ocean. The warmed, buoyant, and deep waters flow toward the surface, causing more deep waters to form.

Superimposed on the large-scale, annual mean flows are seasonal cycles associated with warming or cooling of surface waters. These seasonal cycles are intense in the upper 50 m or so, and have a significant imprint to depths of 100 m, and much more in polar regions. In summer, there is a thin mixed layer at the surface, typically 20–50 m deep, with waters below cooling progressively to annual average values. In wintertime, surface waters are cooler and denser. They thus mix readily with deeper waters. Vertical mixing, and the transfer of waters from the surface to the ocean interior, are thus predominately a wintertime process.

The conflation of mixing processes leaves the oceans with three great realms. There is a warm surface layer extending from about 45° N to 45° S, and from the surface to about 100 m depth. There is a cold water realm (temperature <4°C), extending from the Arctic to Antarctica, and up from the seafloor. The cold water realm includes the surface ocean at latitudes poleward of about 60°. From there to the midlatitudes, its upper boundary progressively deepens to about 1000 m depth, which is typical for the region between 45° N and S. Finally, there is an intermediate realm, which shows a high seasonality at the surface, from ~45° to 60°. As in the polar realm, subsurface waters form at the surface during wintertime, and progressively cooler waters form at progressively higher latitudes. Wintertime surface waters sink to moderate depths to fill the "main thermocline" lying within this intermediate realm and extending from ~200 to 1000 m depth, and from about 60° N to 60° S.

Ocean Biogeochemistry

The interaction between ocean chemistry and biology reflects five generalizations or facts about the ocean. First, all photosynthesis takes place in the sunlit upper layer of the oceans, which is ~100 m deep. Second, almost all metabolism in the oceans is by prokaryotes and single celled eukaryotes. Third, most organic matter is heavier than water and tends to sink. Fourth, almost all organic matter is eventually respired or remineralized (metabolized back to inorganic constituents) rather than preserved in deep-sea sediments. Finally, ocean currents transport dissolved chemicals in the direction of flow, while turbulence mixes waters with higher and lower concentrations and attenuates concentration gradients.

These generalizations explain the basic cycles of biologically active chemicals and their distributions in the oceans. In the upper ocean, single-celled plants (phytoplankton) assimilate carbon, nitrogen, phosphorus, trace metals, and other elements to make tissue. Most of this tissue is rapidly remineralized (transformed to inorganic chemicals by respiration), with the consumption of O_2 and the release of dissolved inorganic carbon, nutrients, and metals. A fraction survives to sink toward the seafloor, and this component is mostly remineralized as it sinks. The process depletes shallow waters in biologically active elements, and enriches subsurface waters. If the oceans were stagnant, nutrients would be completely drained from the sunlit zone and life in the surface would cease. However, upwelling and turbulent mixing return nutrients to the surface, where they are again assimilated by organisms.

The most interesting elements in this process are those in shortest supply relative to the biological demand. These elements, especially nitrogen (N), phosphorous (P), silica (Si), and iron (Fe), are almost completely removed from the sunlit zone over much of the oceans. The exception is in areas where subsurface waters are rapidly upwelling back to the surface. The most important site for this process is in the Southern Ocean, and also in the tropics, though to a much lesser extent. There is a plentiful and steady supply of nutrients in subpolar regions where deep waters mix to the surface during wintertime. In the subtropics, slow diffusion through the thermocline maintains a moderate supply of nutrients to the sunlit waters.

These three sequential processes—assimilation of biologically active elements, sinking of surface waters, and remineralization in the dark ocean—determine the distribution of biologically active elements in the sea. These elements are depleted in surface waters but enriched in deep waters. Remineralization continues along the route of deep water flow, and waters become progressively more enriched in nutrients as they flow from the deep Atlantic to the deep Pacific.

Dissolved inorganic carbon (DIC, the sum of CO_2, HCO_3^- [bicarbonate ion], and CO_3^{2-} [carbonate ion]) in the oceans is obviously utilized by biological activity, but it is never depleted by more that about 10%. There simply is not enough N and P in seawater to support more carbon uptake. Nevertheless, this degree of nutrient utilization has important consequences for atmospheric CO_2. Of the ~40,000 Gt (gigatons) of carbon in "mobile reservoirs" on Earth's surface, about 1.5% is in the atmosphere as CO_2, about 3% is in the land biosphere and soils, and the lion's share is dissolved in the oceans as DIC. The concentration of DIC in seawater, and the pH (or a related property), determine the partial pressure of CO_2 in surface seawater. Since the amount of DIC in the oceans is so much greater than the atmospheric CO_2 inventory, the partial pressure of CO_2 in surface seawater sets the concentration of CO_2 in the atmosphere.

The interplay of three processes can cause changes in the partial pressure of CO_2 (pCO_2) at the sea surface, and in air, over timescales of 10^2–10^4 years. The first process is biological utilization of DIC, which lowers pCO_2. The second is production of skeletal calcium carbonate ($CaCO_3$), which raises pCO_2 by converting 2 HCO_3^- ions into 1 CO_3^{2-} ion and one CO_2 molecule. The third is the riverine input of HCO_3^- and the burial of skeletal $CaCO_3$ in deep-sea sediments. These processes play the fundamental role in glacial-interglacial CO_2 variations.

Over longer timescales other processes dominate, as discussed next.

REGULATION OF ATMOSPHERIC CO_2 AND EARTH'S TEMPERATURE OVER MILLIONS OF YEARS

There is a fairly simple hypothesis for the regulation of Earth's average temperature. It invokes our understanding that volcanism, and other degassing processes associated with Earth's hot interior, steadily add CO_2 to the atmosphere. At the same time, weathering removes CO_2 to balance this input. *Weathering* is the attack of carbonic acid on rock-forming minerals of the solid Earth. In this process, rock-forming minerals are degraded to clay minerals, which are depleted in SiO_2 (silica dioxide) and cations, and CO_2 is converted to HCO_3^-. The dissolved products go into groundwater and eventually to the ocean. An example is the weathering of potassium feldspar to kaolinite:

$$4\,KAlSi_3O_8 + 4\,CO_2 + 6\,H_2O \rightarrow Al_4Si_4O_{10}(OH)_8 + 4\,K^+ + 4\,HCO_3^- + 8\,SiO_2 \qquad (6)$$

If Earth's climate is stable, CO_2 input to the atmosphere by volcanism must be nearly in balance with CO_2 removal by weathering. A simple feedback maintains this balance. If CO_2 input is faster than consumption by weathering, the CO_2 concentration of air rises, and temperature warms. Under these conditions, weathering will accelerate, mainly because chemical reactions speed up as temperature rises. If volcanic input slows, CO_2 will fall, temperatures will cool, weathering will slow, and CO_2 input and output will once again come into balance.

Weathering is very slow below the freezing point, and chemical reactions typically double in rate for a 10° C rise in temperature. Therefore it is possible to compensate for relatively large changes in CO_2 outgassing with relatively modest changes of temperatures.

This idea was originally developed by Walker et al. (1981), formalized into a mathematical model by Berner et al. (1983), and modified and refined by Berner and colleagues in subsequent papers (Berner 2006; Berner and Kothavala 2001). Berner's recent models attempting to explain atmospheric CO_2 changes invoke many other important processes influencing the atmospheric CO_2 balance. Mathematical descriptions of the carbon cycle focus on the past 543 Myr (the Phanerozoic), and we will discuss this work further below.

IMPLICATIONS FOR PALEOCLIMATE

The discussion in this chapter illustrates that there are three reasons that Earth's average temperature might vary. First, the sun could have been shining more or less brightly in the past. Second, the concentration of greenhouse gases could have been higher or lower. Third, Earth's albedo may have changed. The study of paleoclimate involves characterizing Earth's climate in former times, and identifying properties that allow us to distinguish between these three possibilities in order to explain the dynamics of major climate changes of the past.

REFERENCES

Berner, R. (2006), Geocarbsulf: A combined model for Phanerozoic atmospheric O_2 and CO_2, *Geochimica et Cosmochimica Acta*, *70*, 5653–5664.

Berner, R., and Z. Kothavala (2001), Geocarb III: A revised model of atmopsheric CO_2 over Phanerozoic time, *American Journal of Science*, *301*, 182–204.

Berner, R., A. Lasaga, and R. Garrels (1983), The carbonate-silicate geochemical cycle and its effects on atmospheric carbon-dioxide over the past 100 million years, *American Journal of Science*, *260*, 641–683.

Peixoto, J. P., and A. L. Oort (1992), *Physics of Climate*, New York: American Institute of Physics.

Walker, J.C.G., P. B. Hays, and J. F. Kasting (1981), A negative feedback mechanism for the long-term stabilization of Earth's surface temperature, *Journal of Geophysical Research–Atmospheres*, *86*, 9776–9783.