CLIMATE AND THE OCEANS

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Preface

The truth and nothing but the truth, but not
the whole truth.

This is a book on climate, with an emphasis on the role of the ocean. The emphasis is on large-scale processes and phenomena, and on the physical aspects of the ocean rather than its chemical or biological properties. It is not a textbook on physical oceanography, of which there are several good ones, nor is it a textbook on climate, of which there are some good ones. Rather, and as its size may indicate, the book is an introduction to, or a primer on, the ocean–climate system.

This book could be used to provide an introductory "big picture" for more advanced students or for scientists in other fields, or it could provide advanced reading for undergraduate students taking courses at a more elementary level. The book is somewhat more mechanistic than most books at this level: the emphasis is on *how things work,* and in particular how the ocean works and how it influences climate. I discuss observations to motivate the discussion, but the main emphasis is not on what things happen to be, but *why* and *how* they happen to be.

This is a *fast* book, although it is not, I hope, a *loose* book. It covers a lot of ground, quickly, and tries not to get bogged down in too much detail. Having said that, one of the most important questions to answer in the study of climate is to understand just what is a detail and what is essential. If one is studying the climate as a whole, then one might regard the presence of a small island in the northeast Atlantic as a detail, and it surely is (unless one is studying the climate of that island). One might also regard the precise way in which carbon dioxide molecules vibrate and rotate when electromagnetic radiation impinges upon them as a detail, yet it is precisely this motion that gives us the greenhouse effect, which makes our planet habitable, and which also gives us global warming, and which may affect our economy to the tune of trillions of dollars. Hardly a detail! But can we, and need we, understand all such matters to understand climate? Without answering that question, it is clear that in a short book such as this, choices have to be made. I will occasionally, but only occasionally, simply tell you how things are, without delving into the mechanisms if they are tangential to our narrative.

Given that we do move around sharp bends quite quickly, a certain amount of sophistication is assumed of the reader, or at least a willingness to think a little and puzzle things out and perhaps even look up one or two references. On the other hand, I haven't assumed much background knowledge—just a bit of basic physics and mathematics and some general knowledge about the ocean and climate. In other words, I'm writing the book for a smart and motivated but somewhat ignorant reader, and I hope you don't mind being so characterized. In some contrast to most elementary or undergraduate books, I have not shied away from topics that are of current interest or subjects of current research. It should be clear from both the text and the context when the contents of a section are not completely settled or are still a topic of research, rather than being wholly standard textbook material. I have tried to be as objective as possible when discussing such matters, but this attempt does not mean giving two sides of a controversial matter equal weight or suggesting that both are equally valid. Sometimes, an opinion is just plain wrong. The topics of current interest are especially noticeable in the second half of the book, in the chapters and sections on climate variability, global warming, and climate change. The chapter on ocean circulation also reflects our relatively recent understanding of the ocean's meridional overturning circulation. The more mathematical aspects of the book tend to be concentrated in the first half of the book, and the reader who may be surprised that some of the topics dealing with current research are less mathematical should ponder the quotation at the opening of chapter 2. Some readers may wish to skip ahead directly to chapters that particularly interest them, and this way of reading should be possible by referring back to the earlier chapters as needed.

I thank Peter Gent and Carl Wunsch for their perceptive reviews of an early draft of this book. They saved me from myself in a number of ways. Thanks also to

1 BASICS OF CLIMATE

> The climate's delicate, the air most sweet.
>
> —William Shakespeare, *A Winter's Tale*

To appreciate the role of the ocean in climate, we need to have a basic understanding of how the climate system itself works, and that is the purpose of this chapter. Our emphasis here is the role of the atmosphere—we don't pay too much attention to the oceans as we'll get more of that (lots more) in later chapters—and we assume for now that the climate is unchanging. So without further ado, let's begin.

THE PLANET EARTH

Earth is a planet with a radius of about 6,000 km, moving around the sun once a year in an orbit that is almost circular, although not precisely so. Its farthest distance from the sun, or *aphelion*, is about 152 million km, and its closest distance, *perihelion*, is about 147 million km. This ellipticity, or eccentricity, is small, and for most of the rest of the book we will ignore it. (The eccentricity is not in fact constant and varies on timescales of about 100,000 years because of the influence of other planets on Earth's orbit;

these variations may play a role in the ebb and flow of ice ages, but that is a story for another day.) Earth itself rotates around its own axis about once per day, although Earth's rotation axis is not parallel to the axis of rotation of Earth around the sun. Rather, it is at an angle of about 23°, and this is called the *obliquity* of Earth's axis of rotation. (Rather like the eccentricity, the obliquity also varies on long timescales because of the influence of the other planets, although the timescale for obliquity variations is a relatively short 41,000 years.) Unlike the ellipticity, the obliquity is important for today's climate because it is responsible for the seasons, as we will see later in this chapter.

Earth is a little more than two-thirds covered by ocean and a little less than one-third land, with the oceans on average about 4 km deep. Above Earth's surface lies, of course, the atmosphere. Unlike water, which has an almost constant density, the density of the air diminishes steadily with height so that there is no clearly defined top to the atmosphere. About half the mass of the atmosphere is in its lowest 5 km, and about 95% is in its lowest 20 km. However, relative to the ocean, the mass of the atmosphere is tiny: about one-third of one percent of that of the ocean. It is the weight of the atmosphere that produces the atmospheric pressure at the surface, which is about 1,000 hPa (hectopascals), and so 10^5Pa (pascals), corresponding to a weight of 10 metric tons per square meter, or about 15 lb per square inch. In contrast, the pressure at the bottom of the ocean is on average about 4×10^7Pa, corresponding to 4,000 metric tons per square meter or 6,000 lb per square inch!

Table 1.1
Main Constituents of the Atmosphere

Constituent	*Molecular weight*	*Proportion by volume*
Nitrogen, N_2	28.01	78.1%
Oxygen, O_2	32.00	20.9%
Argon, Ar	39.95	0.93%
Water vapor, H_2O	18.02	~0.4% (average) ~1%–4% (at surface)
Carbon dioxide, CO_2	44.01	390 ppm (0.039%)
Neon, Ne	20.18	18.2 ppm
Helium, He	4.00	5.2 ppm
Methane, CH_4	16.04	1.8 ppm

Molecular weight is the molar mass, measured in grams per mole. In addition, there are trace amounts of krypton, hydrogen, nitrous oxide, carbon monoxide, xenon, ozone, chlorofluorocarbons (CFCs), and other gases.

The atmosphere is composed of nitrogen, oxygen, carbon dioxide, water vapor, and a number of other minor constituents, as shown in table 1.1. Most of the constituents are well mixed, meaning that their proportion is virtually constant throughout the atmosphere. The exception is water vapor, as we know from our daily experience: Some days and some regions are much more humid than others, and when the amount of water vapor reaches a critical value, dependent on temperature, the water vapor condenses, clouds form, and rain may fall.

Earth's temperature is, overall, maintained by a balance between incoming radiation from the sun and the radiation emitted by Earth itself, and, slight though it may be compared to the ocean, the atmosphere has a substantial effect on this balance. This effect occurs because water vapor and carbon dioxide (as well as some other minor constituents) are *greenhouse gases*, which means that they absorb the infrared (or longwave) radiation emitted by Earth's surface and act rather like a blanket over the planet, keeping its surface temperature much higher than it would be otherwise and keeping our planet habitable. However, we are getting a little ahead of ourselves—let's slow down and consider in a little more detail Earth's radiation budget.

RADIATIVE BALANCE

Solar radiation received

Above Earth's atmosphere, the amount of radiation, S, passing through a plane normal to the direction of the sun (e.g., the plane in the lower panel of figure 1.1) is about 1,366 W/m^2. (A watt is a joule per second, so this is a *rate* at which energy is arriving.) However, at any given time, half of Earth is pointed away from the sun, so that on average Earth receives much less radiation than this. How much less? Let us first calculate how much radiation Earth receives in total every second. The total amount is S multiplied by the area of a disk that has the same radius as Earth (figure 1.1). If Earth's radius is a, then the area of the disk is $A = \pi a^2$, so that the rate of total radiation received is $S\pi a^2$. Over a 24-hour period, this radiation is spread out over the entire surface of Earth, although not, of course, uniformly. Now, Earth is almost a sphere of

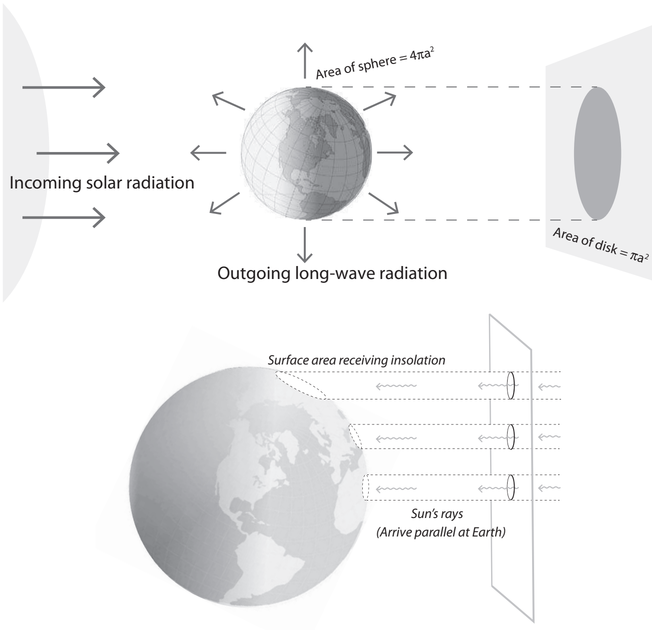


Figure 1.1. Top: The incoming solar radiation impinges on a disk of area πa^2 but is on average spread out over a sphere of area $4\pi a^2$. Bottom: Variation of incoming solar radiation with latitude. A given amount of radiation is spread over a larger area at high latitudes than at low latitudes, so the intensity of the radiation is diminished, and thus high latitudes are colder than low latitudes.

radius a, and the area of a sphere is $4\pi a^2$. Thus, the average amount of radiation that Earth receives per unit area may be calculated as follows:

$$\text{Total radiation received} = S\pi a^2. \tag{1.1}$$

$$\text{Area of Earth} = 4\pi a^2. \tag{1.2}$$

as the *albedo*, of the solar radiation is reflected back to space by clouds, ice, and so forth, so that

$$\text{Net incoming solar radiation} = S_0(1-\alpha) = 239\ \text{Wm}^2, \qquad (1.6)$$

with $\alpha = 0.3$ (we discuss the factors influencing the albedo more below).

This radiation is balanced by the outgoing infrared radiation. Now, of course, Earth is not a blackbody at a uniform temperature, but we can get some idea of what the average temperature on Earth should be by supposing that it is, and so

$$\text{Outgoing infrared radiation} = \sigma T^4. \qquad (1.7)$$

Equating equations 1.6 and 1.7, we have

$$\sigma T^4 = S_0(1-\alpha), \qquad (1.8)$$

and solving for T, we obtain $T = 255$ K or −18°C. For obvious reasons, this temperature is known as the average emitting temperature of Earth, and it would be a decent approximation to the average temperature of Earth's surface if there were no atmosphere. However, it is in fact substantially lower than the average temperature of Earth's surface, which is about 288 K, because of the greenhouse effect of Earth's atmosphere, as we now discuss.

Greenhouse effect

Earth is covered with a blanket of gas made mainly of nitrogen, oxygen, carbon dioxide, and water vapor. This

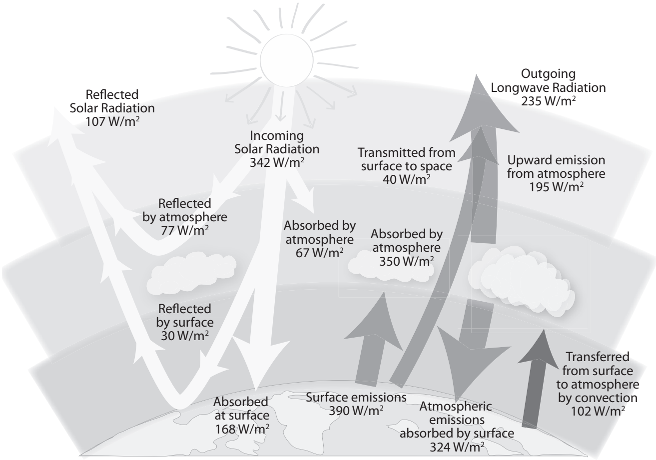


Figure 1.3. The energy budget of Earth's atmosphere, showing the average solar and longwave radiative fluxes per unit area and the convective flux from the surface to the atmosphere. Adapted from Kiehl and Trenberth, 1997.

is higher than this, typically about 0.5, but it can be as high as 0.9 for thick clouds; the albedo of the surface is on average about 0.1 but is much higher if the surface is covered with fresh snow or ice.

The surface is warmed by the solar radiation absorbed and so emits radiation upward, but because the surface temperature of Earth is so much less than that of the sun, the radiation emitted has a much longer wavelength—it is infrared radiation. Now, the atmosphere is *not* transparent to infrared radiation in the same way that it is to solar radiation; rather, it contains greenhouse gases—mainly carbon dioxide and water vapor—that absorb the infrared radiation as it passes through the atmosphere. Naturally enough, this absorption warms the atmosphere, which then re-emits infrared radiation, some of it downward, where it is absorbed at Earth's surface. Thus, and look again at figure 1.3, the total downward radiation at the surface is much larger than it would be if Earth had no atmosphere. Consequently, the surface is much warmer than it would be were there no atmosphere, and this phenomenon is known as the *greenhouse effect*.

A simple mathematical model of the greenhouse effect

Let us now construct a simple mathematical model illustrating the greenhouse effect. Our purpose in doing so is to see somewhat quantitatively, if approximately, whether the atmosphere might warm the surface up to the observed temperature. Let us make the following assumptions:

1. The surface and the atmosphere are each characterized by a single temperature, T_s and T_a, respectively.
2. The atmosphere is completely transparent to solar radiation.
3. Earth's surface is a blackbody.
4. The atmosphere is completely opaque to infrared radiation, and it acts like a blackbody.

The model is illustrated in figure 1.4, which the reader will appreciate is a very idealized version of figure 1.3.

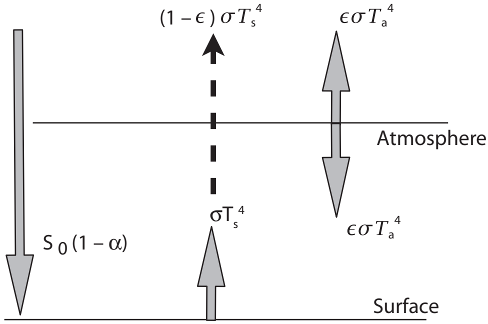


Figure 1.4. An idealized two-level energy-balance model. The surface and the atmosphere are each characterized by a single temperature, T_s and T_a. The atmosphere absorbs most of the infrared radiation emitted by the surface, but it is transparent to solar radiation.

The parameter ϵ, called the emissivity or the absorptivity, determines what fraction of infrared radiation coming from the surface is absorbed by the atmosphere, and we initially assume that $\epsilon = 1$; that is, the atmosphere is a blackbody and absorbs all the surface infrared radiation. The incoming solar radiation, S_0, and the albedo are presumed known, and the unknown temperatures T_s and T_a are obtained by imposing radiative balance at the surface and the atmosphere. At the surface, the incoming solar radiation, $S_0(1 - \alpha)$, plus the downward longwave radiation emitted by the atmosphere is balanced by the longwave radiation emitted by the surface, and therefore

$$S_0(1 - \alpha) + \sigma T_a^4 = \sigma T_s^4. \tag{1.9}$$

2 THE OCEANS: A DESCRIPTIVE OVERVIEW

Persons attempting to find a motive in this narrative will be prosecuted; persons attempting to find a moral in it will be banished; persons attempting to find a plot in it will be shot.

—Mark Twain, *Adventures of Huckleberry Finn*

We now start to look at the ocean(s)[1] in a little more detail, albeit in a rather descriptive manner, as a precursor to the more mechanistic or dynamical description, or "explanation," that we try to provide in chapter 4. That is, in this chapter we describe what's going on but with no underlying organizing principle—with no plot, one might say.

There is a sense in which all explanations are really descriptions; what we may think of as an explanation is really a description at a more general level. Nevertheless, the distinction is useful, at least in science: an explanation does not just describe the phenomenon at hand but also provides some more fundamental reason for its properties, and ideally for the properties of a whole class of phenomena. Descriptions are useful because they are the precursor of explanation, and in this chapter our modest goal is to provide a brief descriptive overview of the oceans and their large-scale circulation, focusing on matters that significantly affect climate.

SOME PHYSICAL CHARACTERISTICS OF THE OCEANS

The ocean basins

The ocean covers about 70% of Earth's surface, and so has a total area of about 3.61×10^{14} km^2. Currently, about two-thirds of Earth's land area is in the Northern Hemisphere, so that about 57% of the ocean is in the Southern Hemisphere, 43% in the Northern; the Northern Hemisphere itself is 61% ocean, and the Southern Hemisphere is about 80% ocean. The ocean's average depth is about 3.7 km, but there are deep trenches where the depth reaches about 10 km. The volume of the ocean is approximately 1.3×10^{18} m^3 and, given that the average density of seawater is about 1.03×10^3 kg/m^3, the total mass of the ocean is about 1.4×10^{21} kg, or 1,400,000,000,000,000,000 metric tons. In comparison, the mass of the atmosphere is about 5×10^{18} kg, about 300 times less: the air at the surface weighs about 1,000 times less than seawater, but the effective vertical extent of the atmosphere is a few times greater than that of the ocean.

The ocean basins have changed over time as the continents have moved and deformed as a consequence of the convection in Earth's mantle that leads to the movement of the tectonic plates and so of the continents themselves, all taking place on a timescale of tens to hundreds of millions of years. The ocean itself has existed for a long time, essentially because the water once formed has nowhere to go: the loss of water vapor to space is negligible because nearly all the water vapor is concentrated at the lowest levels of the atmosphere, and the stratosphere and the upper atmosphere are extremely dry. It is believed that the ocean has in fact existed in roughly the sense that we know it now for perhaps some 3.8 Ga (3.8 billion years), since the beginning of the Archaean era, when Earth cooled sufficiently for land masses to form and water to condense. The water itself had its origins both in volcanism and degassing from Earth's interior, and in collisions with extraterrestrial bodies—probably mainly small icy protoplanets (moonlike bodies) and comets. Such collisions were fortunately much more common in this stage in the evolution of the solar system than they are now. Since that time, the ocean basins have certainly come and gone many times. As continents move significantly on a timescale of tens to hundreds of millions of years, one can envision several distinct continent–ocean configurations over Earth's history, and some of the more recent ones are illustrated in figure 2.1 since the breakup of the "supercontinent" Pangea some 200 million years ago. Reconstruction of the configuration naturally becomes increasingly difficult and so more prone to error the further back one goes in time, but it is believed that there may have been a number of supercontinents over Earth's history, perhaps each a few hundred million years apart.

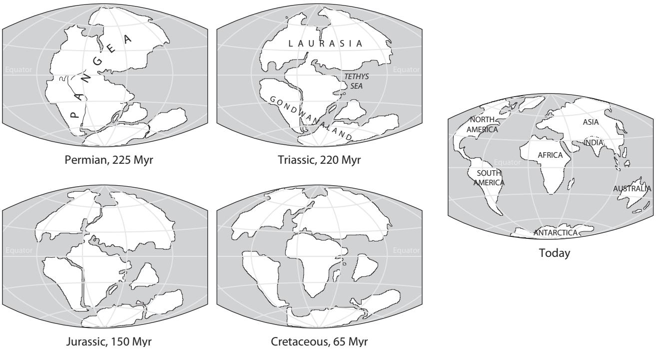
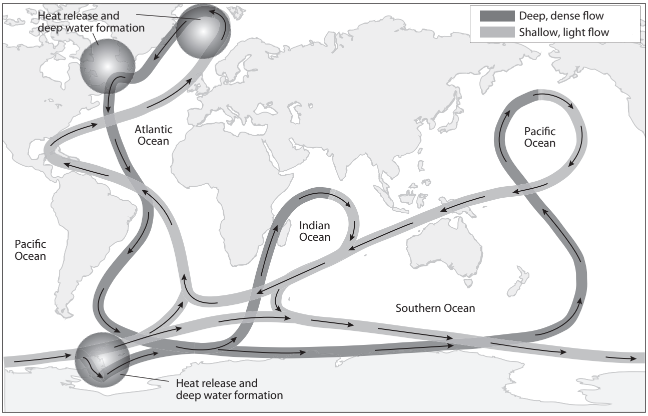


Figure 2.1. Schematic of the configuration of the oceans and continents over the past 225 million years, since the breakup of the supercontinent Pangea. Source: Adapted from USGS (http://pubs.usgs.gov/gip/dynamic/historical.html).

Composition

In today's climate the oceans are mainly liquid; only about 2% of the water on the planet is frozen. Most of the frozen water is in the ice sheets of Antarctica (with 89% of the world's ice, and an average depth of about 2 km) and Greenland (8%, 1.5 km deep). The volume of sea ice, formed by the freezing of seawater, is far less than that of land ice because typically it is only a few meters thick. Its extent also varies considerably by season. However, the importance of land ice and sea ice for climate is in some ways comparable because their areal coverage is similar: about 10% of land is covered with ice year round and about 7% of the ocean. Ice on both land and the ocean has a high albedo, up to 70% when fresh compared to



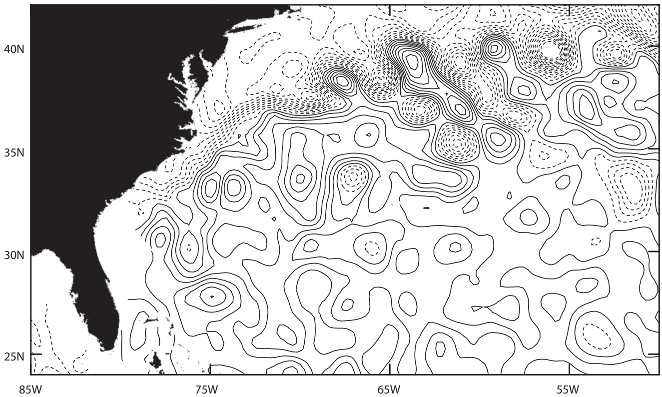


Figure 2.6. Top: An artist's impression of the global ocean circulation, sometimes called the "conveyor belt." Bottom: The sea-surface height in the Atlantic on October 15, 2008, indicating the presence of the Gulf Stream and mesoscale eddies.

two systems ocean eddies tend to be much smaller, with scales of 50–300 km, compared to 500–3,000 km in the atmosphere. In spite of their small scale, the ocean eddies are energetic—the total kinetic energy associated with the eddies in the ocean is about ten times larger than the kinetic energy in the mean currents. Nevertheless, the eddies do not completely destroy the mean flows or make them meaningless, although they can obscure them from easy view and straightforward measurement. In fact, even though eddies dominate the energy budget, they don't dominate the global-scale transport of important properties (such as heat and salinity), and recognizable large-scale oceanic flows remain as a residual after appropriate averaging. Understanding how the eddies and the large-scale mean currents interact remains a daunting challenge in physical oceanography, and the reader may wish to contemplate for a moment the two views of the ocean, juxtaposed in figure 2.6. A full reconciliation of these contrasting views and an understanding of how they fit together are perhaps now coming into our reach, although for now they remain tantalizingly beyond our grasp.

3 A BRIEF INTRODUCTION TO DYNAMICS

Mathematics is the easiest bit in physics.

—Pierre-Gilles de Gennes, *Les Objets Fragile*

We now begin our quest of providing an explanation for how and why the ocean circulates the way it does and how and why it affects the climate. In this chapter, we'll explain some of the basic dynamical principles that determine the circulation; in the next chapter, we'll apply these principles to the circulation itself. The current chapter is a little more mathematical than the others in this book, but it requires no more sophistication on the part of the reader—perhaps less in fact. Nevertheless, some readers may prefer to read the later chapters first, referring back to the material here as needed.

The atmosphere and ocean are both fluids and, although one is a gas and the other a liquid, their motion is determined by similar physical principles and described by similar equations: the *Navier–Stokes equations* of fluid dynamics. These are complex, nonlinear, partial differential equations that require the largest supercomputers to solve, but embedded within them are two important principles that reflect the dominant force balances in the atmosphere and ocean, namely *hydrostatic balance* and *geostrophic balance*. Hydrostatic balance represents the balance in the vertical direction between the pressure force and the gravitational force, and geostrophic balance represents the horizontal balance between the pressure force and the so-called Coriolis force, a force that arises because of the rotation of Earth. If we understand these forces, we will be able to understand a great deal about the motion of the atmosphere and ocean, so let's figure out what they are. We'll begin with the forces that arise (or appear to arise) as a consequence of Earth's rotation, namely centrifugal force and the Coriolis force, and then consider the pressure force. It turns out that for most geophysical applications, the Coriolis force is much more important than centrifugal force, but we need to understand the latter to understand the former, so that is where we begin.

CENTRIFUGAL FORCE

Suppose that you are riding in a train that starts to go around a bend rather quickly. You feel like you are being thrust outward toward the side of the car, and if you are really going quickly around a tight curve, you might have to hang onto something to stay put. The outward force that you are feeling is commonly known as *centrifugal force*. Strictly speaking, it is not a force at all (we'll explain that cryptic comment later), but it certainly feels like one. What is going on?

One of the most fundamental laws of physics, Newton's first law, says that, unless acted upon by a force, a body will remain at rest or continue moving in a straight line at a constant speed. That is, to change either direction or speed, a body must be acted upon by a force. Thus, in order for you to go around a bend, a force must act (and act on the train too), and this force, whatever it may be in a particular situation, is called the *centripetal force*. Without that force, you would continue to go in a straight line. The centrifugal force that you feel is caused by your inertia giving you a tendency to try to go in a straight line when your environment is undergoing a circular motion, so you feel that you are being pushed outward. You do end up going around the bend because your seat pushes against you, providing a real force (the aforementioned centripetal force) that accelerates you around the bend. The centripetal force that makes the train go around the bend comes from the rails pushing on the train wheels. With this discussion in mind, we see that there are two ways to think about the force balance as you go around the bend (literally).

1. From the point of view of someone standing by the side of the tracks, you are changing direction and a real force is causing you and the train to change direction, in accord with Newton's laws.
2. From your own point of view, you are stationary relative to the train. If you don't look out of the window, you don't know you are going around a bend. There appear to be two forces acting on you: the centrifugal force pushing you out and the seat pushing back (the centripetal force) in the other direction. In this

frame of reference, Newton's law is satisfied because the two forces cancel each other out. That is, in the train's frame of reference you remain stationary and there is a balance of forces between the centripetal force from the seat pushing you in and the centrifugal force pushing you out.

The centrifugal force is, therefore, really a device that enables us to use Newton's laws in a rotating frame of reference: we can say that Newton's laws are satisfied in a rotating frame provided we introduce an additional force, the centrifugal force. There is a simple formula for this force that is derived in appendix A to this chapter. If an object of mass *m* is going around in a circle of radius *r* with a speed *v*, then the centrifugal force, F_{cen}, is given by

$$F_{cen} = \frac{mv^2}{r}. \tag{3.1}$$

The centrifugal force per unit mass is just v^2/r.

A quite analogous situation occurs for us on Earth. Earth is actually rotating quite quickly; it goes around once a day, and the velocity of Earth's surface at the equator is a quite respectable 460 m s^{-1}, or a little faster than 1,000 miles per hour. Sitting on the surface of Earth, we must therefore experience a centrifugal force that is trying to fling us off into space. The reason that we stay comfortably on the surface is that the force of gravity overwhelms the centrifugal force, as a quick calculation shows. The radius of Earth is about 6,300 km, so that using equation 3.1, the centrifugal force per unit mass is given by

$$\frac{460^2}{6.3 \times 10^6} = 0.03 \text{ m s}^{-2}. \tag{3.2}$$

This value should be compared to the force of gravity per unit mass (i.e., the gravitational acceleration) at Earth's surface, which is 9.8 m s^{-2}. The centrifugal effect is therefore quite small, although not so small that we cannot measure it. In fact, because the centrifugal effect is largest at the equator and diminishes to zero at the poles, over time Earth has developed a slight bulge at the equator such that the line of apparent gravity (the real gravity plus the centrifugal force) is perpendicular to Earth's surface. (The distance from the center of Earth to the surface is in fact some 30 km larger at the equator than at the poles.) The centrifugal force is otherwise not terribly important, and its effect can be taken into account by slightly modifying the value of the gravitational force as necessary.

THE CORIOLIS FORCE

Another apparent force is caused by Earth's rotation, one that only arises when bodies are in motion relative to the rotating Earth, and this force is known as the *Coriolis force* after the French engineer and scientist Gaspard-Gustave Coriolis (1792–1843). It turns out to be much more important than the centrifugal force for currents and winds, although its effects are rather subtle. The sphericity of Earth is not important in the Coriolis force itself, and until we get to the section on differential rotation and Earth's sphericity in this chapter, we can imagine Earth to be a rotating disk, with the rotation axis through the North Pole at the center of the disk.

To gain an intuitive idea of what the coriolis force is, consider the (hopefully fanciful) situation in which a missile is launched from the North Pole toward the equator, as illustrated in figure 3.1.[1] Once launched and above Earth's atmosphere, the missile is uninfluenced by the fact that Earth is rotating beneath it. Suppose the missile is initially aimed at Africa and that it takes about six hours for the missile to reach the equator. When the missile reaches the equator, Earth has rotated a quarter turn, but the missile has not and so it will land in South America! From the perspective of someone tracking the missile from the surface of Earth, the missile has not gone in a straight line but has veered to the right.

If a missile is fired from the equator toward the North Pole, it begins its flight with a large eastward velocity, equal to that of the surface of Earth at the equator. As the missile moves poleward, it maintains an eastward velocity (in fact, it conserves its angular momentum), which soon exceeds that of Earth beneath it. From the point of view of an observer on Earth's surface, the missile again appears to veer to the right. From the point of view of an observer on Earth, it seems that the missile has experienced a force—the Coriolis force—that causes it to veer to the right. Just as with the centrifugal force, the Coriolis force is something that we introduce to be able to use Newton's laws in a rotating frame of reference. It is not a real force in the sense that no other body causes it;

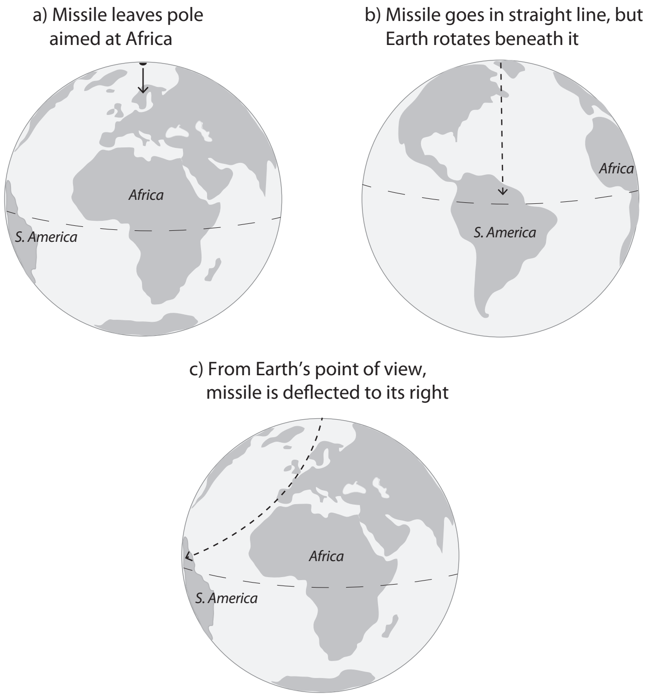


Figure 3.1. A missile launched from the North Pole toward Africa. Earth rotates beneath the missile, and the missile lands in South America. From the point of view of an observer on Earth, the missile has been deflected to its right, and the force causing that deflection is the Coriolis force.

rather, it is a manifestation of the inertial tendency of a body to go in a straight line while Earth rotates.

Let us now imagine that a missile is fired along a line of latitude, neither toward the equator nor away from it. First consider the situation in which the missile is fired

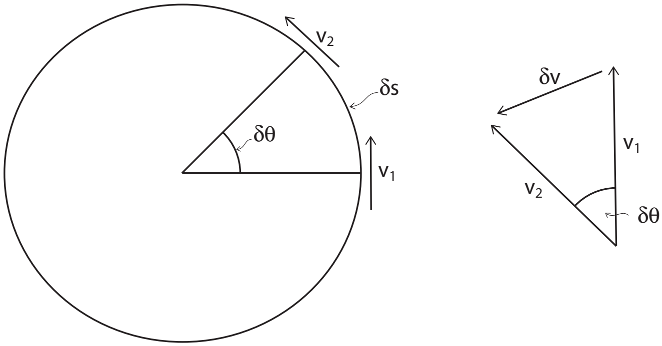


Figure 3.5. A body moving in a circle is constantly changing its direction and so accelerating. The acceleration is directed toward the center of the circle and has magnitude v^2/r, where v is the speed of the body and r is the radius of the circle.

$v = r\Omega = r\,\delta\theta/\delta t$, but its direction is changing such that at all times, the direction of motion is perpendicular to the radius; thus, if at the initial time the position vector is $\boldsymbol{r}_1$, then its velocity, $\boldsymbol{v}_1$, is in the perpendicular direction, and a short time later the velocity $\boldsymbol{v}_2$ is perpendicular to its new position vector, $\boldsymbol{r}_2$.

As can be seen in figure 3.5, the angle through which the radius vector has moved is related to the distance moved by

$$\delta\theta = \frac{\delta s}{r} \approx \frac{|\delta r|}{r}. \tag{3.13}$$

Because the velocity is always perpendicular to the radius, the direction of the velocity must move through the same angle as the radius and we have

$$\delta\theta = \frac{|\delta \boldsymbol{v}|}{v}. \tag{3.14}$$

The acceleration of the body is equal to the rate of change of the velocity; using equation 3.14, we obtain

$$\frac{\delta|\boldsymbol{v}|}{\delta t} = v\frac{\delta\theta}{\delta t} = \frac{v^2}{r}, \tag{3.15}$$

where the last equality uses the fact that $v = r\delta\theta/\delta t$. That is to say, the acceleration of a body undergoing uniform circular motion is directed along the radius of the circle and toward its center and has magnitude

$$a_{\mathrm{circ}} = \frac{v^2}{r} = \Omega^2 r. \tag{3.16}$$

This kind of acceleration is called *centripetal acceleration*, and the associated force causing the acceleration (for there must be a force) is called the centripetal force and is directed inward, toward the axis of rotation.

Forces in the rotating frame of reference

Now let us consider the motion from the point of view of an observer undergoing uniform circular motion. An illuminating case to consider is that of a satellite or space station in orbit around Earth; when an astronaut goes for a space walk from the space station, she appears to be weightless, with no forces whatsoever acting upon her (figure 3.6). Now in fact, the astronaut is undergoing circular motion around Earth, and is therefore accelerating, and the force providing the acceleration is Earth's gravity.

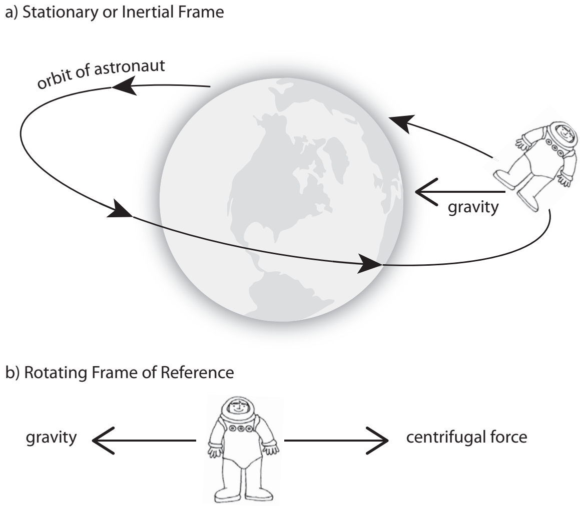


Figure 3.6. An astronaut orbiting Earth. Panel a views the motion in a stationary frame of reference, in which Earth's gravitational force provides the centripetal force that causes the astronaut to orbit Earth. Panel b views the situation from the astronaut's frame of reference, in which the gravitational force is exactly balanced by the centrifugal force and the astronaut feels weightless.

But from the point of view of an observer rotating with the astronaut, the astronaut is stationary and therefore no net forces are acting upon her. Now we know that gravity is acting, pulling her toward Earth, and we say that this force is balanced by another force, *centrifugal force*, which is pushing her out. The forces exactly balance, so the astronaut appears weightless; that is, in the rotating frame, the centripetal gravitational force pulling her toward Earth is exactly balanced by the centrifugal force pushing her out. Because the centripetal force has magnitude $\Omega^2 r$, the centrifugal force must have this magnitude also, and we conclude that in a rotating frame of reference, there appears to be an additional force, the centrifugal force, which acts to accelerate a body along a radius, outward from the axis of rotation. The magnitude of the centrifugal force is $\Omega^2 r$ per unit mass, where r is the distance from the axis of rotation.

There is no centrifugal force in inertial frames of reference. However, given that we live on a rotating planet, it is extremely convenient to describe motions on Earth from a rotating frame of reference in which Earth's surface is stationary, and in this frame it appears that there is a centrifugal force that is trying to fling us into space. Rather fortunately for those of us living on the planet's surface, the centrifugal force is much weaker than Earth's gravity.

Coriolis force

In this section, we give a mathematical, although elementary, derivation for the magnitude of the Coriolis force.

Coriolis force for a body moving zonally

Consider a disk rotating with an angular velocity Ω, so that at a radius r from the axis of rotation the disk has a tangential velocity $u_d = \Omega r$. Suppose that an observer is sitting on the disk and is thus stationary in the disk's

4 THE OCEAN CIRCULATION

This is a court of law, not a court of justice.

—Oliver Wendell Holmes

The climate in general and the oceans in particular are complicated systems, and if one is not careful it is easy to lose sight of the forest for the trees. For that reason, a useful philosophy is to begin with an austere picture of the phenomenon at hand and then gradually add layers of complexity and detail. The first picture will be a simplification, but if it is based on sound scientific principles, then it will provide a solid foundation for what follows, and it will become possible to work toward an understanding of the system as it really is. In this chapter we apply this philosophy to try to understand the ocean circulation. We won't seek a full understanding of the real system; rather, we will construct a physical and mathematical representation of it, a model based on the same laws of physics that are satisfied by the real ocean.

WHAT MAKES THE OCEAN CIRCULATE?

As we discussed in chapter 2, it is useful to think of the large-scale ocean circulation as having two main components:

a quasi-horizontal circulation consisting of the gyres and other surface-enhanced currents, and a deeper overturning circulation, the meridional overturning circulation. What makes the ocean go around this way? What "drives" the ocean, if anything? Bypassing the ambiguous term "drive," there are three main distinct physical phenomena that lead to the circulation of the ocean:[1]

1. The mechanical force of the wind on the surface of the ocean provides a stress that produces a quasi-horizontal circulation that includes, most noticeably, the *wind-driven gyres*. The predominantly horizontal currents of the world's ocean, shown in figure 2.3 in chapter 2, are primarily a consequence of wind forcing. Less obviously, the wind also plays a role in producing a deep, interhemispheric meridional overturning circulation, a circulation in which the water sinks near one pole and rises near the other.
2. Buoyancy effects, caused mainly by the cooling of the oceans at high latitudes and heating at low latitudes, generally produce denser water at high latitudes. Salinity is a secondary source of density gradients in today's climate. An overturning circulation arises in response to these density gradients with cool, dense water sinking at high latitudes, moving equatorward and rising at lower latitudes and/or in the opposite hemisphere.
3. The mixing of fluid properties, and in particular heat, by small-scale turbulent motions (sometimes called turbulent diffusion) brings heat down into the abyss and enables an overturning circulation to be maintained.

The gyres and other quasi-horizontal currents are mainly a response to winds, and although they are affected by buoyancy effects and mixing, we can safely call them wind driven. The meridional overturning circulation (MOC), on the other hand, involves all three effects in an essential way. Most obviously, the MOC arises as a response to the surface density gradients (item 2 in our list) and is sometimes called the *thermohaline circulation*, so-called because it is enabled by the buoyancy effects of heat and salt leading to the sinking of dense water. However, we will see that the MOC can only be *maintained* if either mixing or wind is present, for they enable the deep water to rise to the surface to begin circulating anew. Without them, the deep circulation would stagnate.

Let's first discuss the wind-driven circulation, the great gyres, and western intensification, and follow that with a discussion of the MOC. The equatorial currents are different again, and we defer discussing them until chapter 6.

THE WIND-DRIVEN CIRCULATION AND THE GREAT GYRES

To better understand how the processes described above produce an ocean circulation like that described in chapter 2, let us consider an idealized ocean, with much simplified geometry, and see if we can first understand how that works. Our idealized view of the gyres is illustrated in figure 4.1, which the reader will appreciate is an abstraction

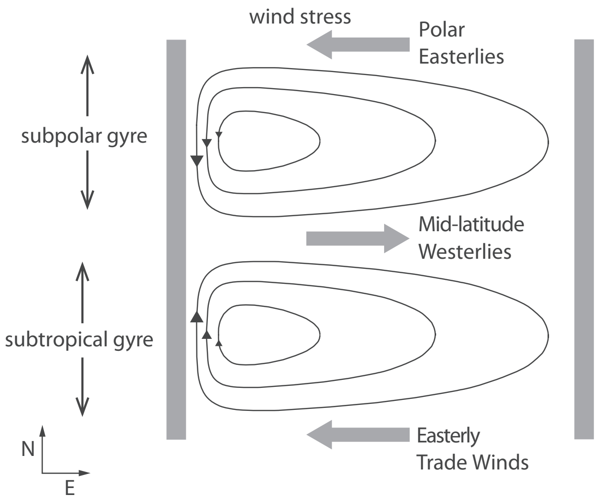


Figure 4.1. An idealized gyre circulation in a rectangular ocean basin in the Northern Hemisphere, showing the subtropical gyre (lower, typically extending from about 15°N to 45°N), the subpolar gyre (upper), and the intense western boundary currents on the left.

of the real circulation of the world's ocean. The main questions we wish to answer are relatively simple:

1. Why do the gyres exist in the first place? What determines the way they go around and how strong they are?
2. Why are they more intense on the western sides of the oceans?

The gyres exist because the mean winds provide a mechanical forcing, a stress, on the oceans, and this stress causes the water to accelerate. For the oceans to be in mechanical balance, the imposed forces must be counteracted by frictional forces where the water rubs against the ocean bottom or side. Frictional forces only arise when the water is in motion, so that if there is a wind blowing, then the ocean must be in motion, and an overall balance between the wind and the frictional forces ultimately comes about. However, there are important effects caused by Earth's rotation that determine the structure of the gyres, as we will see.

For the sake of definiteness, we consider the subtropical gyre in a rectangular ocean—the lower gyre of figure 4.1. The winds blow eastward on the poleward side of the gyre (these are the midlatitude westerly winds) and westward at low latitude (the tropical trade winds), and it seems entirely reasonable that the ocean should respond by circulating in the manner shown. However, in the last chapter we noted that Earth's rotation plays a significant role in large-scale circulation and that flows are generally in geostrophic balance, except for the Ekman layer in the upper ocean, where the flow is at right angles to the wind. How does this description square with the notion of a gyre that seems to go around in the same direction as the wind?

The Ekman and wind-induced geostrophic flows

We show first that the wind does indeed induce a geostrophic flow that has the same sense as the wind itself. The mean winds are to the east in midlatitudes and to the west in the tropics and, as we showed in the section in chapter 3 on Ekman layers, there is a flow in the upper

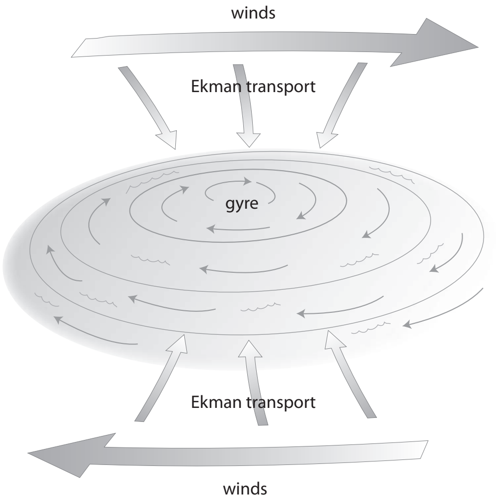


Figure 4.2. Production of gyres by winds. The winds blowing as shown induce a converging Ekman flow, causing the sea level to increase in the center, thus giving rise to a pressure gradient. This gradient in turn induces a geostrophic flow around the gyre, in the same sense as the winds themselves.

ocean at right angles to the wind. As illustrated in figure 4.2, this combination causes the flow to converge in the center of the gyre. This convergence pushes up the surface of the ocean, causing the sea surface to form a gentle dome, with the ocean surface at the center of the gyre a few tens of centimeters higher than at the edges. The converging fluid must go somewhere, and the only place for it to go is downward. A complementary situation arises in the subpolar gyre, where the westerly (eastward) winds displaced, then there is a net upward force on it, and the object moves upward until it floats on the surface. If the solid object is heavier than the fluid displaced, the object sinks. These considerations apply to water itself. If we cool the water at the surface of the ocean, or add salt to it, it becomes more dense and therefore sinks—and it can sink quite quickly. A parcel of water that is negatively buoyant at the surface of the polar ocean can sink to considerable depth in a concentrated convective plume in a matter of hours to days, with a corresponding vertical velocity of a few centimeters per second. Similarly, if we warm the water that is at the bottom of the ocean, it will become lighter and rise, although this tends to be a much slower process, spread out over a wide area.

The overturning circulation maintained by mixing

How do the above considerations apply to the circulation of the ocean? For simplicity, we consider only the effects of temperature and not of salinity, and a schema of the circulation is given in the top panel of figure 4.6. The ocean, is, roughly speaking, a big basin of water for which the temperature of air just above the sea surface decreases with latitude. Air–sea exchange of heat heats or cools the water at the sea surface so that it has, approximately and on average, the temperature of the air above it. The sea-surface temperature thus decreases more or less monotonically from the equator to the pole, and as a consequence the density of the water at the sea surface increases from the equator to the pole.

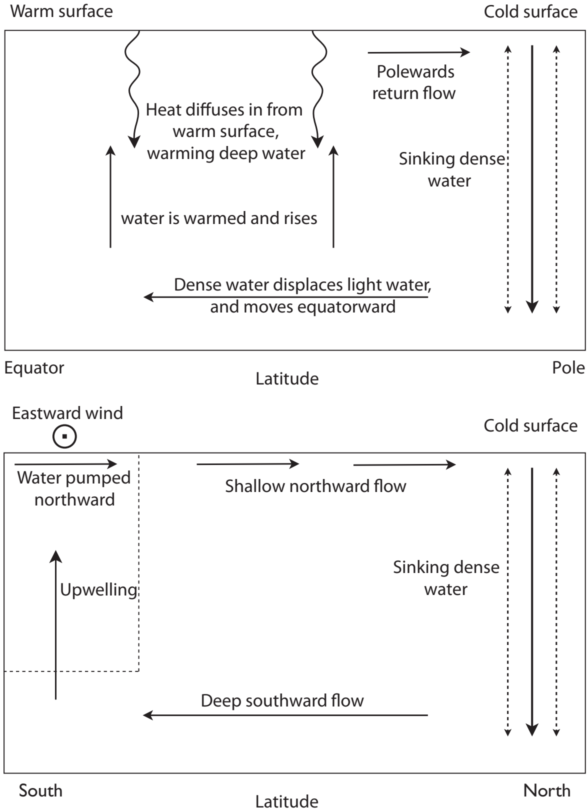


Figure 4.6. Schema of the two main components of the MOC. Top: The mixing-maintained circulation. Dense water at high latitudes sinks and moves equatorward, displacing warmer, lighter water. The cold, deep water is slowly warmed by diffusive heat transfer (mixing) from the surface in mid- and low latitudes, enabling it to rise and maintain a circulation. Bottom: Winds over the Antarctic Circumpolar Current (outlined by dashed lines) pump water northward, and this pumping enables deep water to rise and maintain the circulation. In the absence of both wind and mixing, the abyss would fill up with the densest available water and the circulation would cease.

As we mentioned, a fluid parcel itself sinks if it is cold and sufficiently dense. This is just what happens to water at high latitudes, especially in winter in the North Atlantic and near Antarctica, and this process is known as convection. Some lighter water at depth comes up to the surface to take the place of the dense, sinking water, as indicated by the dashed lines in figure 4.6, and as this water comes into contact with the cold atmosphere, it too cools and sinks, so that eventually the whole column of water at high latitudes is cold and dense. What happens then? Recall that the pressure at some level in a fluid is equal to the weight of the fluid above that level, so that if a column of fluid is cold and therefore dense, then the column weighs more than does a column of lighter fluid. Thus, the pressure in the deep ocean is largest at the high latitudes because the cold water weighs more than the warmer water at low latitudes. Thus, in the deep ocean there is a pressure force acting to push fluid from high latitudes to low latitudes, and the water begins to circulate, flowing at depth from high latitudes to low latitudes.

If no other physical processes occurred, the dense water would displace light water until the entire deep ocean were filled up with cold, dense water with polar origins. Nearer the surface, there would be a region of strong vertical temperature gradients, linking the low temperature of the abyss with the warmer surface waters. However, the deep, abyssal waters would eventually stop circulating because the water in the deep ocean would be as cold and dense as the coldest and densest waters at high latitudes at the surface. That is, the surface water would no longer be denser than the water beneath it, and convection and the deep circulation would cease. This state would be the "cold death" of the ocean.

So what enables a deep circulation to continue? The circulation continues because the deep water in low and midlatitudes is continually, albeit weakly, warmed by the transport of heat from the surface. This warming enables the water to rise and the circulation to continue. If there were no such heat transport, the deep ocean would simply fill up with cold, dense polar water. There would then be no convection because the cold surface waters at high latitudes would not be negatively buoyant. Thus, although the circulation can be thought of as being set up by a buoyancy gradient at the surface, its continuation relies on the effect of transport of heat down into the abyss, and without that, this part of the overturning circulation could not be maintained.

What physical process causes the downward heat transfer? In a *quiescent* fluid, the heat is transferred by molecular diffusion, in which molecules of water pass on their energy to neighboring molecules without any wholesale transport of fluid itself. However, the molecular diffusivity is very small and molecular diffusion is a slow process indeed, requiring thousands of years for a significant amount of heat to be diffused from the surface to the abyss. In fact, the ocean is a turbulent fluid, and the downward transport of heat is mainly effected by small-scale turbulent eddies. This process is sometimes called turbulent diffusion because the process is similar to that of molecular diffusion but with parcels of water replacing individual molecules. (Turbulent diffusion arises in large part from internal gravity waves that break and mix the fluid. Such waves, analogous to waves on the surface of the ocean but interior to the fluid, are generated by mechanical forcing—by the winds and the tides. Thus, without the effects of mechanical forcing, this component of the MOC would be weak indeed because the diffusion would be small.) Thus, to summarize, the following two effects combine to give an overturning circulation.

1. A meridional buoyancy gradient between the equator and the pole enables dense water to form at the surface at high latitudes and then potentially to sink in convective plumes and move equatorward. In today's ocean, the buoyancy gradient is predominantly produced by the temperature gradient.
2. The slow warming of the abyssal waters by turbulent diffusion of heat from the surface in mid- and low latitudes imparts buoyancy to the deep water and enables it to rise. Without this warming, the abyss would fill with cold, dense water and circulation would cease.

It is natural to think of the meridional buoyancy gradient as being between the equator and the pole, mainly caused by temperature falling with latitude. In this case, we can envision a meridional circulation in each hemisphere, with sinking at each pole and rising motion in mid- and low latitudes, in both hemispheres.

If one hemisphere were to be significantly colder than the other, then the abyss in both hemispheres could be expected to fill up with the water from the colder and denser hemisphere, which would create an interhemispheric circulation (more on that later). Finally, although we've couched our description in terms of the temperature effects on buoyancy, the effects of salinity can also be important. Salty water is heavier than freshwater at the same temperature, so adding salt can have a similar effect to that of cooling the surface. In today's climate, temperature has a larger effect than salinity on the variations in buoyancy so that the circulation is thermally driven, rather than salt driven. However, variations in salinity turn out to be the key difference in the overturning circulation of the Atlantic and the Pacific—the North Atlantic is saltier than the North Pacific, and so it can more easily maintain an overturning circulation.

The overturning circulation maintained by wind

The second mechanism that can lead to a deep overturning circulation relies, in its simplest form, on the presence of strong zonal wind blowing over the ocean surrounding Antarctica, as illustrated in the lower panel of figure 4.6. Unlike an ocean basin, the ocean surrounding Antarctica is effectively a channel, for it has no meridional boundaries and so no real gyres. Let's first look at the flow in this channel, and then look at how this flow affects the global overturning circulation. The wind around Antarctica blows in a predominantly zonal direction, toward the east. As one might expect, the wind generates a mean current in the same direction—the Antarctic Circumpolar Current, or ACC. However, because Earth is rotating, the wind stress generates an Ekman flux (as described in chapter 3) that is perpendicular to the wind, and so northward (the Coriolis force deflects bodies to the lef*t* in the Southern Hemisphere), as illustrated in figure 4.7.

The northward-flowing water in the Ekman layer must be compensated by southward-moving water to maintain a mass balance. In a gyre, the return flow could be at the surface in a western boundary current, but none exist in the ACC and the flow must therefore return at depth, where friction along the bottom enables the flow to be nongeostrophic, or the presence of topography allows zonal pressure gradients to be maintained. Where does the deep water ultimately come from? One option would be that the flow simply circulates locally in the Southern Hemisphere. However, if the water in the Northern Hemisphere is sufficiently dense, then it will be drawn into the Southern Hemisphere and into and across the ACC, where it can come up to the surface. Water at high latitudes in the North Atlantic is in fact sufficiently dense for this to occur, although water in the North Pacific is not (the key difference is salinity—the North Atlantic is saltier than the North Pacific). Thus, the presence of winds in the Southern Ocean generates an interhemispheric meridional overturning circulation, in which water sinks at high northern latitudes

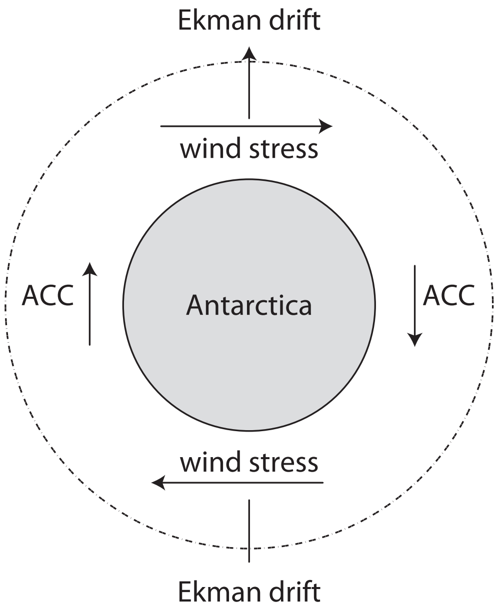


Figure 4.7. Schematic of the flow in the Antarctic Circumpolar Current (ACC). The wind predominantly blows in a zonal direction around the Antarctic continent, generating an Ekman flow toward the north and a net loss of water from the channel. The water returns at depth, generating a deep overturning circulation, as illustrated in the bottom panel of figure 4.6.

and moves southward across the equator, upwelling in the ACC. Unlike the mixing-maintained circulation described in the previous section, no mixing is required to draw up the deep water; rather, the wind itself pumps the deep water up.

Putting it all together

Thus, to summarize, the meridional overturning circulation has two mechanistically distinct components: a component maintained by mixing and a component maintained by wind, both responding to the surface buoyancy distribution. The two can exist side by side, and the overturning circulation in the Atlantic Ocean is schematically illustrated in figure 4.8. Some of the water that sinks in the North Atlantic moves across into the Southern Hemisphere and upwells in the ACC (enabled by the wind), and some upwells and returns in the North Atlantic itself (enabled by mixing). The water that sinks in the North Atlantic (forming the North Atlantic Deep Water) does not in fact extend all the way to the bottom of the ocean because there is some even denser water beneath it—Antarctic Bottom Water, which comes from high southern latitudes and circulates through the effects of mixing.

Which component of the circulation is dominant? Only careful observations can tell us, although currently it is often believed that the wind component is stronger than the mixing component in the Atlantic Ocean. The North Pacific Ocean is generally less dense than the North Atlantic because it is fresher; also it does not support a vigorous interhemispheric circulation and so partakes more weakly in the global-scale overturning circulation that is sketched in figure 2.6. Note finally that the horizontal velocities in the abyssal ocean are usually quite small, on the order of 1 mm s^{-1}, and at this speed it would take some 300

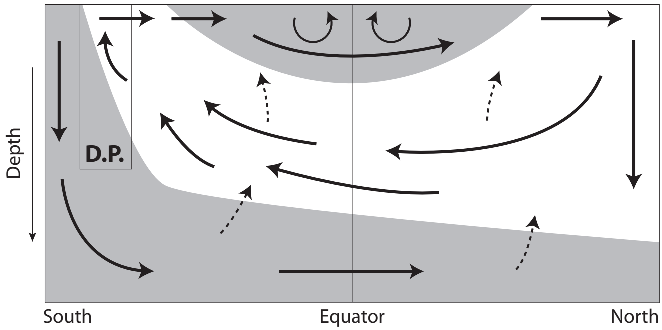


Figure 4.8. Schematic of the meridional overturning circulation, most applicable to the Atlantic Ocean (D.P. indicates the Drake Passage, the narrowest part of the ACC). The arrows indicate water flow, and dashed lines signify water crossing constant-density surfaces, made possible by mixing. The upper shaded area is the warm water sphere, including the subtropical thermocline and mixed layer, and the lower shaded region is Antarctic Bottom Water. The bulk of the unshaded region in between is North Atlantic Deep Water.

years for a parcel to move from its high-latitude source to the equator, still longer if the path were not direct. Thus, if the surface conditions change, it will take several hundred years for the deep ocean to re-equilibrate.

OCEAN CIRCULATION IN A NUTSHELL

The large-scale ocean circulation may usefully be divided into a quasi-horizontal circulation, comprising the gyres and other surface and near-surface currents, and a meridional overturning circulation. Embedded within the circulation are smaller mesoscale eddies, which actually contain the bulk of the kinetic energy of the ocean and which are analogous to atmospheric weather systems.

The ocean gyres

- The ocean gyres are primarily wind driven, responding in particular to the north–south variations of the zonal wind. The subtropical gyres lie between about 15° and 45° in both hemispheres, with the subpolar gyres poleward of that in the Northern Hemisphere.
- The wind stress has a direct effect in the uppermost few tens of meters of the ocean, where it induces an Ekman flow at right angles to the wind. This Ekman flow in turn causes the sea surface to slope and produces a geostrophic flow, which is the main component of the gyres and which extends down several hundred meters.
- The main gyres all have a strong intense current at their western boundary (e.g., the Gulf Stream in the North Atlantic, the Kuroshio in the North Pacific), which arises from the combined effects of Earth's sphericity and its rotation.

The overturning circulation

- The overturning circulation is a response to variations in surface buoyancy, in that the densest water at the surface (usually at high latitudes) sinks and moves away from the sinking region at depth.
- For the circulation to persist, the deep water must be brought up to the surface; otherwise, the abyss will fill up with the densest water available and then stagnate. Two processes bring deep water up to the surface: mixing and the wind.

(*continued*)

— Mixing warms the deep water at low latitudes, which may then rise through the thermocline, maintaining a circulation of sinking at high latitudes and rising at low latitudes.
— Strong westerly winds in the Antarctic Circumpolar Current can draw water up from the deep and induce an interhemispheric circulation, which is particularly strong in the Atlantic.

The other main currents

- The Antarctic Circumpolar Current is the collection of eastward flowing currents around Antarctica, which taken together form the largest sustained current system on the planet. It is a response both to wind and to the meridional temperature gradient.
- The equatorial current systems are predominantly controlled by the winds, consisting typically of a westward flowing current and eastward countercurrents and undercurrents.

APPENDIX A: MATHEMATICS OF INTERIOR FLOW IN GYRES

Suppose that the wind blows zonally across the ocean, with a stronger eastward wind to the north, as in figure 4.3. Away from coastal regions (where friction may be important) the forces present are the zonal wind force (which here we simply denote F_w^x), the Coriolis force (fv and fu) and the pressure gradient force ($\partial\phi/\partial x$ and $\partial\phi/\partial y$, where $\phi = p/\rho$), and we represent their balance mathematically as

$$-fv = -\frac{\partial \phi}{\partial x} + F_w^x, \qquad fu = -\frac{\partial \phi}{\partial y}, \tag{4.5a, b}$$

in the zonal and meridional directions, respectively. If there were no wind, the flow would be in geostrophic balance, and indeed the flow is in geostrophic balance at depths greater than 100 m or so, below the level at which the winds' effects are directly felt. Conservation of mass also gives a relation between u and v, namely

$$\frac{\partial u}{\partial x} + \frac{\partial v}{\partial y} = 0. \tag{4.6}$$

If we cross-differentiate equation 4.5 (i.e., differentiate equation 4.5a with respect to x and equation 4.5b with respect to y and subtract), then the divergence terms vanish using equation 4.6 and the pressure gradient terms cancel, and we obtain

$$\beta v = -\frac{\partial F_w^x}{\partial y}, \tag{4.7}$$

where $\beta = \partial f/\partial y$ is the rate at which the Coriolis parameter increases northward. The balance between the varying wind and the meridional flow embodied in equation 4.7 is known as Sverdrup balance, and the effect of differential rotation is called the beta effect. If the wind stress has a positive curl, that is, if $\partial F_w^x/\partial y > 0$, then, because β is also positive, v must be negative and the interior flow must be equatorward. There must be a poleward return flow in a boundary current at either the western or the eastern edge of the ocean basin, where the effects of friction conceivably can be such as to balance the Coriolis and wind stress curl terms. But only if the flow returns in the western boundary current can the frictional effects balance the wind stress curl overall, for then the flow overall has the same sense as the wind.

5 THE OCEAN'S OVERALL ROLE IN CLIMATE

> The coldest winter I ever spent was
> a summer in San Francisco.
>
> —Mark Twain

The ocean plays a number of roles in our present climate, and in this chapter we discuss two of the most important:

1. The ocean moderates the climate by taking in heat when the overlying atmosphere is hot, storing that energy and releasing heat when the atmosphere is cold.
2. The ocean redistributes heat in the large-scale ocean circulation.

In addition, the ocean generally has a lower albedo than land, so that if all the ocean were replaced by land, the planet as a whole would be cooler. In some contrast, when the ocean freezes it forms sea ice, which has a generally high albedo. Thus, if the climate as a whole were to warm up, then the sea-ice extent would likely diminish, lowering the overall albedo and so further warming the planet. And finally, of course, the ocean is far and away the main reservoir of water on the planet, and if the planet were dry the atmosphere would have no clouds and the greenhouse effect would be much smaller, with wholesale changes in the climate. These last few effects are a little indirect, so let's focus on the two effects we mentioned first.

THE MODERATING INFLUENCE OF THE OCEAN

Perhaps the most obvious effect that the ocean has on climate is its moderating effect on extremes of temperature, both diurnally (i.e., the day–night contrast) and annually (the seasonal cycle). We focus on the effects on the annual cycle because these tend to be on a larger scale and more befitting a book with *climate* in the title, but much the same principles and effects apply to the diurnal cycle. First we take a look at the observations to confirm that there *is* a moderating influence from the ocean. Fig. 5.1 shows the annual cycle of temperatures of San Francisco and New York. The two cities have similar latitudes (San Francisco is at about 38° N and New York is at about 41° N) and both are on the coast, yet we see from the figure that the range is enormously larger in New York—the highs are higher and the lows are lower. (One wonders if the respective climate extremes affect or even effect the different personalities of New Yorkers and Californians.)

The difference is mainly caused by the fact that the climate of San Francisco is *maritime*, meaning that it

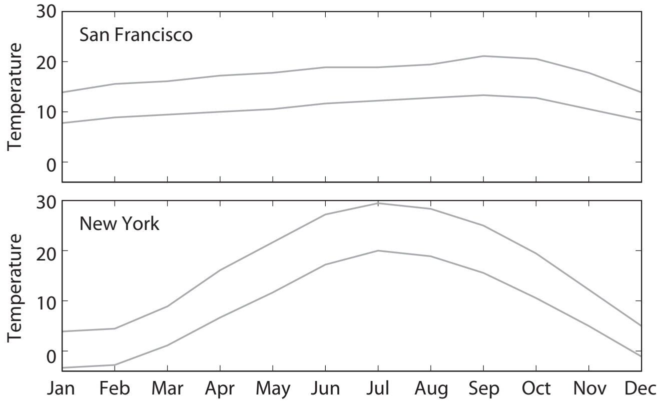


Figure 5.1. The seasonal cycle of temperature (°C) in San Francisco and New York. For each city, we plot the average low temperature and the average high temperature for each month. Note the much bigger range in New York and the maximum earlier in the year, in July rather than September.

is influenced by the ocean, whereas the climate of New York is, in spite of it being on the Atlantic coast, essentially continental. A city that is truly land-locked, such as Moscow, has a climate much more like New York than San Francisco. New York's climate is continental because the mean winds come primarily from the west, so they blow over land and take up its temperatures before they reach the city. In contrast, the winds have blown over the Pacific Ocean before arriving at San Francisco. So why does the ocean moderate the climate? It is in part because water has a relatively high heat capacity, compared to the material that makes land (e.g., soil and concrete), and in larger part because the upper ocean is in

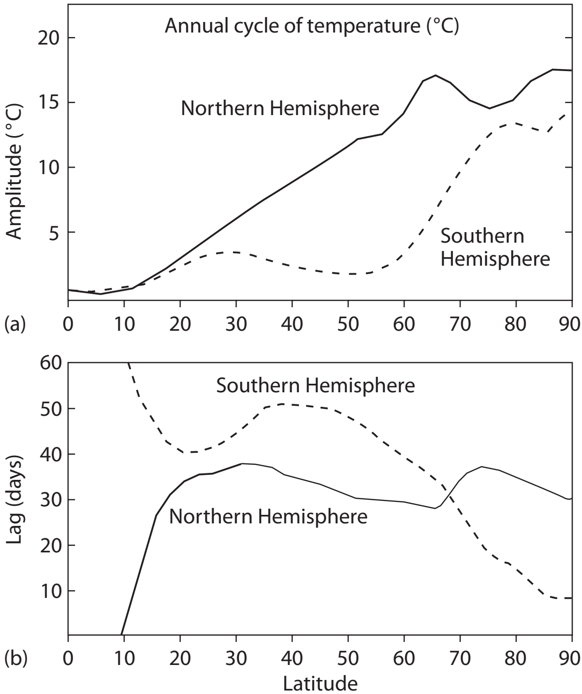


Figure 5.2. Amplitude and lag of the annual cycle in the Northern and Southern hemispheres, as a function of latitude. The lag is the time, in days, from the maximum solar insolation to the maximum temperature. Source: Trenberth, 1983.

The lag in the seasons

The observant reader noted in figure 5.1 that not only is the seasonal cycle more muted in San Francisco, but also that the maximum temperatures occur later in the season, in September. This again is an effect of the large heat capacity of the system, as a simple argument shows. Suppose that a system is heated externally (e.g., by the sun) and is cooled by the effects of longwave radiation and that the cooling is proportional to the temperature itself. If the system has a very small heat capacity, then the heating and cooling must balance each other at all times. A consequence of this is that the cooling is greatest when the heating is greatest, and so the temperature itself is highest when the sun is highest in the sky. Indeed, we find that in continental climates the temperature is highest fairly soon after the summer solstice and coldest soon after the winter solstice: In Fig. 5.1, we see that New York is hottest in July and coldest in January.

If a system has a large heat capacity, it takes some time to warm up and cool down, and so the maximum temperatures occur some time after the maximum insolation and thus later in the summer. The same effect occurs on a daily basis: inland, the maximum daily temperature occurs shortly after noon, whereas at the seaside the maximum temperature is later in the afternoon. On a large scale, in the Northern Hemisphere midlatitudes, the maximum temperature occurs on average about 30 days after the maximum solar insolation, whereas in the more maritime Southern Hemisphere, the maximum occurs about 45 days after peak insolation (figure 5.2). At very high latitudes, where the Southern Hemisphere is covered by land (Antarctica) but the Northern Hemisphere by ocean (the Arctic Ocean), the lag is longer in the Northern Hemisphere. A mathematical demonstration of this effect is given in appendix A of this chapter.

The general damping of climate variability by the ocean

Not only does the ocean provide a moderating influence on the march of the seasons, but it also can provide a moderating influence on the variability of climate on other timescales too. We talk more about the mechanisms that give rise to climate variability in the next chapter, but for now let us just suppose that the climate system excluding the ocean is able to vary on multiple timescales, from days to years. Then, just as the ocean is able to damp the seasonal variability, the ocean damps variability on all these timescales. However, the ocean does not damp the variations equally on all timescales; rather, because on longer timescales the ocean itself can heat up or cool down in response to climate variations, the damping effects are larger on shorter timescales. We give a brief mathematical treatment of this argument in the next section, and a more complete treatment in appendix A of this chapter.

Mathematical treatment of damping

The surface temperature of the ocean and the land are maintained by a balance between heating and cooling. The heating occurs both by solar radiation and by downward longwave radiation from the atmosphere and is proximately independent of the temperature of the surface itself. The cooling, on the other hand, increases with the temperature—a hot object cools down faster than a warm one. If for simplicity we suppose that the cooling rate varies linearly with temperature, then we can model the surface temperature by the equation

$$C\frac{\mathrm{d}T}{\mathrm{d}t} = S - \lambda T. \tag{5.1}$$

Here, S is the heating source, T is the temperature, and t, the time. The parameter C is the heat capacity of the system, and λ is a constant that determines how fast the body cools when it is hot. Obviously, this equation is too simple to realistically describe how the surface temperature varies (it ignores lateral variations, for one thing), but it illustrates the point we wish to make.

The equation says that the heat capacity times the rate of the temperature increase (the left-hand side) is equal to the heating (S) minus the cooling (λT). If we set $S = 0$ for the moment, then a solution of this equation is $T = T_0 \exp(-\lambda t/C)$, where T_0 is the initial temperature. If S is a constant, then the full solution is

$$T = \frac{S}{\lambda} + T_0 \exp(-\lambda t/C). \tag{5.2}$$

This equation tells us that if there is a perturbation to the system, the perturbation will decay away on the timescale C/λ. With a mixed-layer depth of 100 m and $\lambda = 15\ \mathrm{Wm}^{-2}\ K^{-1}$ (which is suggested by observations for air–sea interactions), we obtain a timescale of a little less than a year. That is to say, the ocean mixed layer can absorb or give out heat on the timescale of about a year. Variability on timescales significantly longer than this is not greatly damped by the presence of an ocean mixed layer because on these timescales the mixed layer itself heats up and cools down and so provides no damping to the system. However, on timescales much shorter than this, the mixed layer absorbs heat from a warm atmosphere, or alternatively gives up heat to a cold atmosphere, thus damping the variability that the atmosphere otherwise might have. The land surface has a much smaller heat capacity, so that the timescale C/λ is much smaller for land than it is for the ocean. There is thus a much smaller damping effect over land than over the ocean.

The situation is not *quite* as straightforward as this argument suggests. A complicating factor is that the entirety of the ocean mixed layer does not respond to fast variations in the atmosphere. Thus, for example, only the top few meters of water may respond to diurnal variations in temperature, and such variations are therefore damped less than one might expect. Nevertheless, the overall effect is clear: The heat capacity of the ocean mixed layer damps variations on timescales up to and including the annual variations. Interested readers can find a more complete description of this effect in appendix A of this chapter.

OCEAN HEAT TRANSPORT

The other great effect that the oceans have on the mean climate is that they transport heat, usually poleward, thus cooling the tropics and subtropics and warming high latitudes. Let's first look at how much the transport is, then we'll discuss the ocean processes that give rise to the transport, and finally what effects the transport has on climate.

How much?

On average, both the atmosphere and the ocean transport heat poleward, and this transport is illustrated in figure 5.3. The total transport of the atmosphere plus the ocean may be determined fairly directly from satellite measurements. Over the whole planet, there is a balance between the incoming solar radiation and outgoing longwave radiation, and if there were no heat transport, the incoming solar radiation would equal the outgoing infrared radiation at each latitude—a state of pure radiative balance. In fact, at low latitudes there is an excess of incoming solar radiation, whereas at high latitudes there is an excess of outgoing infrared radiation, meaning that at low (high) latitudes Earth is colder (warmer) than it would be if it were in pure radiative balance. The imbalance arises because heat is transported poleward by the motion of the atmosphere and ocean, and if we measure the imbalance at each latitude, then we obtain the total heat transport by the atmosphere and ocean. Perhaps needless to say, this measurement is easier said than done, but the advent of modern satellites that make separate measurements of solar and infrared radiation makes it possible. The most accurate estimates come from the period of the Earth Radiation Budget Experiment, in particular over the period 1985–1989, when intense observations were made, but data continue to be gathered.

The Gulf Stream and the climate of Britain and Ireland

It is occasionally said that Britain and Ireland owe their mild climate to the presence of the Gulf Stream and the North Atlantic Drift. Thus, the story goes, the Gulf Stream brings warm water from Florida up the eastern seaboard of the United States and then across the Atlantic in the North Atlantic Drift to the shores of Britain and Ireland, hence moderating the otherwise cold winters. Certainly, the surface temperature of the eastern North Atlantic is a few degrees warmer than the water at the same latitude off the coast of Newfoundland, as figure 2.2 in chapter 2 illustrates. Although this difference does have some effect on the temperature differences between the two locations, Britain and Ireland have a moderate winter climate primarily as a consequence of the fact that they are next to the ocean, with the ocean on their west. Even if there were no gyres in the ocean at all, the climate of these parts would be much more moderate than the climate at similar latitudes on the eastern sides of continental land masses. Thus, Britain and Ireland have a much more similar climate to British Columbia, at a similar latitude on the west coast of Canada, than they do to Newfoundland and Labrador on the east coast. The effects of the east–west asymmetry of sea-surface temperatures on the seasonal climate of Britain and Ireland, and of midlatitude coastal areas surrounding the ocean basins generally, are relatively small.[1]

However, the effects of the ocean and the ocean circulation on the climate of Britain and Ireland are far from small. If the ocean were to cease circulating altogether—that is, both the gyres and the meridional overturning circulation were to cease—then the high latitudes would generally get colder, as we discussed in the previous section, and possibly freeze over. If the oceans did not freeze, western Europe would still have a maritime climate and a more moderate seasonal cycle than the eastern United States and eastern Canada.

APPENDIX A: THE MATHEMATICS OF THE RELATIONSHIP BETWEEN HEATING AND TEMPERATURE

In this appendix, we give an elementary mathematical treatment of the relationship between heating and temperature. We will explain two things: why the temperature range is smaller if a body has a larger heat capacity and why there is a lag between heating and temperatures.

We model the system with the simple equation

$$C\frac{\mathrm{d}T}{\mathrm{d}t} = S - \lambda T. \tag{5.3}$$

Here, S is the heating source, T is the temperature, and t, the time. The parameter C is the heat capacity of the system, and λ is a constant that determines how fast the body cools when it is hot. The equation says that the heat capacity times the rate of the temperature increase (the left-hand side) is equal to the heating (S) minus the cooling (λT). Let us further suppose that the heating is cyclic, with $S = S_0 \cos \omega t$, where ω is the frequency of the heating and S_0 is its amplitude.

To solve the equation, we write $S = \mathrm{Re}\, S_0 \exp(\mathrm{i}\omega t)$ and seek solutions of the form $T = \mathrm{Re}\, T_0 \exp(\mathrm{i}\omega t)$ where Re means "take the real part" and T_0 is a constant to be determined. Substituting into equation 5.3, we have

$$CT_0 \mathrm{i}\omega \mathrm{e}^{\mathrm{i}\omega t} = S_0 \mathrm{e}^{\mathrm{i}\omega t} - \lambda T_0 \mathrm{e}^{\mathrm{i}\omega t}, \tag{5.4}$$

where only the real part of the equation is relevant. From this equation, we straightforwardly obtain

$$\begin{aligned} T &= \mathrm{Re}\,\frac{S_0(\lambda - \mathrm{i}C\omega)\,\mathrm{e}^{\mathrm{i}\omega t}}{\lambda^2 + C^2\omega^2} \\ &= S_0 \frac{\lambda \cos \omega t + \omega C \sin \omega t}{\lambda^2 + C^2\omega^2}. \end{aligned} \tag{5.5}$$

What does this equation tell us? If the heat capacity is negligible, or if the frequency ω is very small (i.e., very slow variations in forcing), then

$$T \approx \frac{S_0}{\lambda} \cos \omega t. \tag{5.6}$$

The temperature is in phase with the heating, and the amplitude of the cycle is S_0/λ. If the heat capacity is large or the frequency is high, then

$$T \approx \frac{S_0}{C\omega} \sin \omega t. \tag{5.7}$$

6 CLIMATE VARIABILITY FROM WEEKS TO YEARS

Climate is in the eye of the beholder.

In this chapter we look at climate variability, and in particular climate variability that is associated in one way or another with the ocean. This condition is not very restrictive because nearly all forms of climate variability on timescales of months to decades are affected by, or even caused by, the ocean. Even in cases in which the underlying cause of the variability is nonoceanic, the ocean may modulate the variability and determine its timescale, and in many ways we can think of the ocean as the pacemaker of climate. We won't talk about climate variability on timescales of centuries to millennia and longer, for that deserves a book on its own. Rather, our focus will be on midlatitude climate variability on intraseasonal and interannual timescales, and on climate variability associated with El Niño. Before delving into all that, let's first discuss if and how climate is different from weather.

CLIMATE AND WEATHER

What is the difference between climate and weather? It is intuitively clear what the weather is—it is the day-to-day state of the atmosphere at some location, usually with particular reference to such things as temperature, windiness, and precipitation. It is also intuitively clear that when we speak of climate we wish to average out all these day-to-day fluctuations and refer to some kind of average of the weather. But what precisely? There is no ideal definition of climate, but a useful working notion is that climate is the statistics of the weather—the mean, the standard deviation, and so forth. However, this notion slightly begs the question of how we calculate the statistics—how do we take the mean, for example? And if climate is a time average of the weather, then how can climate have any temporal variability?

Although it is rather fanciful, it is useful to envision a thought experiment in which we take the *ensemble average* of the weather. Thus, we envision a large number of identical planet Earths, forced the same way, but each one started out in a slightly different way so that each has different weather. We could then unambiguously define the climate to be the average, along with other relevant statistical quantities like the variance, over the ensemble of planet Earths. If the forcing were to change, perhaps because the CO_2 levels in the planets were to increase, then the climate of the ensemble would also change.

The problem with this definition is that it is not practical, there is no such ensemble in reality. However, we can try to take our average in such a way that it mimics the ensemble average as closely as possible, and this way of proceeding will be useful to the extent that the

However, the ocean does have dynamics of its own. The gyres contain smaller eddies that produce considerable variability, just as the atmosphere has weather, and the large-scale gyres themselves may vary on interannual to decadal timescales. The meridional overturning circulation also varies, and its sluggish nature suggests that the variability may include timescales of decades and longer. If this variability is able to produce large-scale, long-lived SST anomalies, it is possible that these anomalies may, over time, affect the average behavior of the atmosphere. Even though the impact of mid- and high-latitude SST anomalies on the atmosphere seems to be small, if the anomalies persist for long enough, they will have a cumulative effect. Such an effect may be responsible for the seeming persistence of the NAO pattern in certain decades that we alluded to previously, but the evidence is not definitive.

EL NIÑO AND THE SOUTHERN OSCILLATION

In the next few sections we describe the phenomenon (and what a phenomenon it is!) known as *El Niño,* or sometimes as El Niño and the Southern Oscillation (ENSO). The Southern Oscillation is the atmospheric part of the phenomenon, and El Niño, the oceanic component; ENSO is the combination. However, unless we need to be particularly precise, we often just use the term El Niño, for this has a pleasant euphony lacking in the acronym ENSO.[2] El Niño is the largest and most important phenomenon in global climate variability on interannual timescales. In brief, it is an anomalous warming of the surface waters in the eastern equatorial Pacific, most pronounced just off the coast of Peru but extending westward to the dateline.

In the next few sections we describe in a little more detail what El Niño is, what mechanisms give rise to it, and what effects it has, or is perceived to have, on weather and climate throughout the world.

What is El Niño?

Every few years the temperature of the surface waters in the eastern tropical Pacific rises quite significantly. The strongest warming takes place between about 5° S and 5° N, and from the west coast of Peru (a longitude of about 80° W) almost to the dateline, at 180° W, as illustrated in figure 6.3. The warming is significant, with a difference in temperature up to 6°C from an El Niño year to a non–El Niño year. The warmings occur rather irregularly, but typically the interval between warmings ranges from three to seven years, as illustrated in figure 6.4.

The warmings have become known as El Niño events, or even (with a little violence to the Spanish language) El Niños. The name derives from the fact that the warm waters off Peru appear at about Christmastime, and the name el Niño is Spanish for the Christ child.[3] The warmings typically last for up to a year, sometimes two, and appear as an enhancement to the seasonal cycle, with high temperatures appearing at a time when the waters are already warming. Although there is no universally

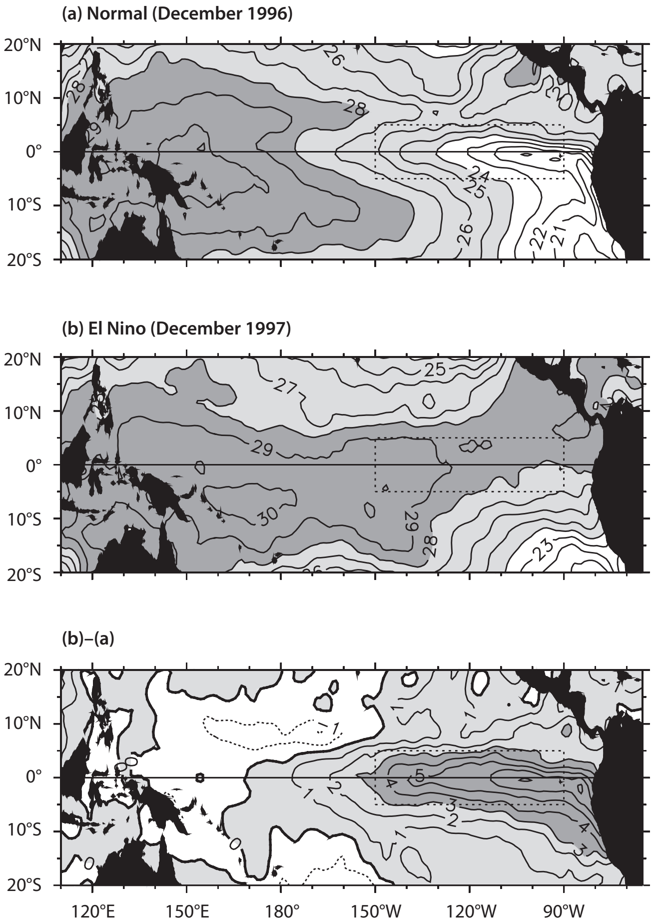


Figure 6.3. The sea-surface temperature in December of a normal (i.e., non–El Niño) year (December 1996, top panel); in a strong El Niño year (December 1997, middle panel); and their difference (bottom panel). A normal year is characterized by a cold tongue of water in the eastern tropical Pacific, which disappears in El Niño years.[5]

as *La Niña* events (la niña, without capitalization, is Spanish for young girl; there is no female equivalent of the Christ child, el Niño in Spanish). We have direct observational evidence—that is, compilations of measurements of the sea-surface temperature from ships and buoys—of El Niño events for more than a century, but the events have almost certainly gone on for a much longer time, perhaps millennia or longer. We know this through a variety of proxy records; tree rings provide some of the most detailed information. As we will discuss later on, El Niño events bring anomalous rainfall and temperature throughout the equatorial Pacific and western North America, affecting the tree-ring characteristics and providing convincing evidence of the occurrence of El Niño events for the past several hundred years.

Corals also provide a good record of El Niño events because they have skeletal growth bands that, rather like tree rings, provide an accurate annual chronology. The isotopic composition of oxygen contained in the skeletons responds to both SST and rainfall, thus providing a record of El Niño that goes far into the past. Indeed, one recent analysis of corals from Papua New Guinea suggests that El Niño events have occurred for the past 130,000 years—that is, even over the last ice age![7]

Our knowledge of El Niño seems to have begun with observations of the SST off the coast of Peru, but it is now understood that the phenomenon itself is Pacific-wide and also involves the atmosphere. Observations of the winds and the surface pressure in the equatorial Pacific show that these tend to covary with the SST. In particular, during El Niño events the equatorial Pacific trade winds, which normally blow toward the west, become much weaker and may even reverse. One useful measure is the pressure difference between Darwin, at the far north of Australia (12° S, 130° E), and Tahiti (17° S, 150° W), an island in the Pacific; the record of this difference is known as the *Southern Oscillation*. (There is nothing truly special about these locations vis-à-vis El Niño, but pressure measurements have been made there for a long time.) We see in figure 6.4 that the Southern Oscillation and the SST record of El Niño are highly correlated.

The mechanism of El Niño

The mean state of the atmosphere and the ocean

Before discussing what processes conspire to produce El Niño events, let us discuss what the mean state of the atmosphere and ocean are, beginning with the atmosphere. The trade winds throughout the equatorial region blow predominantly from higher latitudes toward the equator, and from the east to the west. The low-level convergence at the equator forces the air to rise and then, several kilometers above the surface, move poleward, sinking in the subtropics at about 30° north and south and then moving equatorward at the surface. The meridional cell is known as the Hadley cell, and if there were no continents, the equatorial convergence would be on average the same at all longitudes.

EL NIÑO IN A NUTSHELL

1. El Niño refers to the warming of the surface waters in the eastern equatorial Pacific. The interval between warm events is typically from two to seven years but is quite irregular and can be longer.
2. El Niño events are associated with a weakening of the trade winds and a shifting eastward of the region of convection. The atmospheric side of the events is known as the Southern Oscillation, and the entire phenomenon is known as the El Niño–Southern Oscillation, or ENSO.
3. In contrast to an El Niño event, from time to time the far western equatorial Pacific becomes warmer than usual and the eastern Pacific, colder, which is known as a La Niña event.

What causes El Niño events?

1. El Niño is caused by the mutual interaction between the atmosphere and ocean in the equatorial Pacific, involving a positive feedback between the sea-surface temperatures and the strength of the trade winds.
2. The trade winds normally blow westward, but during an El Niño event the trade winds weaken, allowing the temperatures in the eastern equatorial Pacific to rise, further weakening the trade winds and so on.

What are some consequences of El Niño?

1. The global surface temperature in an El Niño year is up to 0.5°C higher than normal.
2. Convective rainfall in North Australia and Indonesia is suppressed in El Niño years, whereas rainfall is enhanced in western tropical South America.

3. During strong El Niño events, the atmospheric storm track in the eastern Pacific is stronger and further south than normal, bringing heavy rain to central and southern California and penetrating inland across the southern United States.
4. The Atlantic storm track may also be affected, moving south and bringing conditions resembling a negative NAO index.
5. El Niño years are associated with the suppression of Atlantic hurricanes.

is almost reversed, with warmer and wetter weather in the Indonesian archipelago and northern Australia and generally cooler and dryer weather in coastal Peru and Ecuador.

Distant effects

By distant effects we mean the effects of El Niño on other regions of the globe that are not in themselves part of the ENSO cycle. Some of these effects are noticeable, especially if the El Niño event is strong, but others are weak and ambiguous and only emerge after averaging over a number of El Niño events to remove the natural variability—or noise—that causes one year to have different weather from another in any case. We first note the overall effect: In an El Niño year, the globally averaged surface temperature can be as much as 0.5°C higher than the years before and after, and this increase is accounted for by the fact that the surface temperatures in the eastern tropical Pacific can be several degrees higher than normal.

One effect that certainly arises during strong El Niño events is that the storm track over the eastern North Pacific becomes stronger, wetter, and further south in winter, bringing heavy rain to central and southern California and additional snow to the southern Sierras. In both the 1982–83 and 1997–98 events, extensive flooding and landslides occurred. In some contrast, further north in Oregon the snow pack was less than usual because of a warm winter but no enhanced precipitation, and Canada tends to experience warmer and drier winters in El Niño years. These effects are not nearly as pronounced in years with weaker El Niño events, and it is often hard to definitively attribute anomalous winter rainfall to an El Niño event.

The storm track does not end when it reaches California, and New Mexico, Arizona, and indeed much of the southern United States can receive enhanced precipitation in El Niño years. (Indeed, New Mexico has some of the best tree-ring records of El Niño.) The storm track may pick up again over the Atlantic and bring enhanced precipitation to Europe. Certainly, there is a correlation between El Niño events and the phase of the North Atlantic Oscillation, with warm events associated with a negative phase of the NAO. Elsewhere in the world, the effects are weaker, but there is some evidence that parts of East Africa (particularly Kenya and Tanzania) experience wetter weather during warm events.

Finally, we mention another tropical effect, a nonlocal one but one that seems to be quite robust. It is that

Atlantic hurricanes tend to be suppressed during El Niño years. Why should this be? Hurricanes form over warm tropical waters, and when the SST in the Atlantic is anomalously high, then the Atlantic hurricane season is more active than usual. However, hurricanes do not respond to the sea-surface temperature alone. Rather, they are a kind of heat engine, and a significant factor in driving hurricanes is the temperature *difference* between the sea surface and the upper atmosphere. (The effect is analogous to the fact that midlatitude weather systems respond to the temperature gradient between the equator and the poles.) During El Niño events, the Pacific region becomes particularly warm, and this warmth spreads throughout the tropics in the upper atmosphere, and in particular the upper tropical atmosphere over the Atlantic becomes anomalously warm. Hence, the temperature difference between the surface and the upper atmosphere that drives Atlantic hurricanes is reduced, and the Atlantic hurricane season in an El Niño year tends to be less active than usual.

7 GLOBAL WARMING AND THE OCEAN

> Plurality should not be posited without necessity.
>
> —Occam's Razor (attributed to William of Ockham)

In this final chapter we discuss a topic of great current and likely future interest, namely *global warming*. In the first half of the chapter, we talk about warming quite generally: what it is, what the evidence is for it, what the consequences might be, and what the level of uncertainty might be about future warming. In the second half of the chapter, we concentrate on the role and effects of and on the ocean. We find that the evidence unambiguously points to a single culprit for global warming, namely the increasing concentration of greenhouse gases in the atmosphere. Although it is possible that we could, with some effort, come up with other explanations that would fit the facts, we would need to invoke several different explanations to fit all the facts and/or invoke rather implausible ones.

GLOBAL WARMING ITSELF

Global warming, in the sense that it is commonly used, is the observed increase in the average temperature of Earth's surface and atmosphere since the late nineteenth century and its projected continuation. The fact that warming has occurred over the past century and is continuing is undeniable. The degree to which it will continue in the future, and the causes of that warming, are both topics of considerable scientific and societal interest. Global warming is, of course, a form of climate change, albeit a forced one and not a natural one. The name "global warming" is useful because it crisply evokes the global nature of the issue: nowhere will be unaffected, nothing will be impervious, no one will be immune. But of course the problem cannot be properly encapsulated by a single global number. The effects will be worse in some places than others, there may be floods here and droughts there, and these regional changes will dictate how society responds or fails to respond to it. Nevertheless, many effects will scale with the change in the globally averaged temperature, so let us initially focus on that and describe the warming that has occurred over the past century or so.

The observed global temperature record

Since the late nineteenth century, the observed average surface temperature has been increasing, as shown in figure 7.1. The data from land comes from about 4,000 stations distributed widely over the the globe, although naturally enough there are more stations in North America and Europe and fewer stations in such places as Antarctica, Greenland, Siberia, and the Sahara Desert. The actual measurements typically are taken twice daily

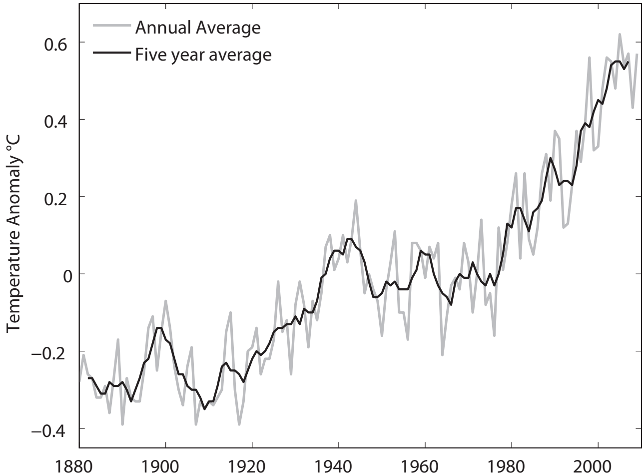


Figure 7.1. The instrumental record of global average surface temperatures from 1880 to 2009, relative to the mean temperature from 1951 to 1980.

(more frequently in some locations) and are of the air temperature a few meters above ground. The ocean data mostly come from in situ observations from ships and buoys; the measurements are the temperature of seawater itself, although in fact this is a good surrogate for the air temperatures just above the surface.[1]

As the temperature data on both the land and the ocean are quite nonuniform, the temperatures are first interpolated onto a regular grid from which a globally averaged temperature can be constructed. (Measurements have in fact been taken on land and on sea since well before 1850, but not with sufficient spatial density to enable the direct construction of a gridded data set.) Obviously, such are found to be small, contributing an error of about 0.005°C per decade, and about 0.05°C over the past century, although it should be said that some critics of the temperature record believe this error estimate to be too small. Satellite measurements, discussed below, provide another check on urbanization errors.

A somewhat different source of error, although still a bias error and one that has certain similarities with the urbanization problem, is that the way that temperatures have been measured has changed over the past century, both over water and on land. On land the early shelters were fairly heterogeneous; they have been slowly replaced with more standardized shelters known as *Stevenson screens* (after Thomas Stevenson, 1818–1887). These shelters are essentially ventilated white boxes that shield the thermometer against precipitation and direct radiation but that allow air to flow past the thermometer and so give accurate measurements of air temperature. Still more recently, some of these shelters have been replaced with mechanically aspirated shelters to further increase the airflow. The errors due to the different box designs are generally thought to be very small (< 0.1°C) at any given station up to 1950 and negligible after that.

Over water, the methods of taking sea-surface temperature have also varied over time, from taking samples in wooden and then canvas buckets in the early part of the record to measuring the temperatures of the water coming into the engine rooms of ships for cooling. Studies suggest that the errors are small, certainly compared to the changes in temperature over the past century, although changes in measuring techniques have led to some apparently artifactual small jumps in the record.[2]

Satellite measurements

Given that there are possible errors in the direct measurements of temperature, it is useful to compare them with satellite measurements, which provide a completely independent record, less influenced by urbanization issues. Of course, satellites have not been taking measurements for as long as the surface record has existed, and they too are subject to their own errors—difficulties both in calibration and in accounting for the fact that the orbit of a satellite tends to decay over time, potentially contaminating the results.[3] Nevertheless, the combination of satellite and surface observations gives a rather powerful check on temperature increase of the past few decades, as shown in figure 7.2.

Satellite measurements are taken with a microwave sounder, which measures the microwave radiation in several bands, as well as an infrared sounder, which makes similar measurements in the infrared band. The brightness in each band is sensitive to both the temperature and the amount of water vapor in the atmosphere, but by measuring in multiple bands a temperature profile of the atmosphere can be constructed. The temperature trends measured by the satellites agree well with those directly measured at the surface and with those in the lower atmosphere measured by *radiosondes* (instruments carried on weather balloons). The surface measurements show a trend of 0.16°C per decade over the past three decades, and the satellite

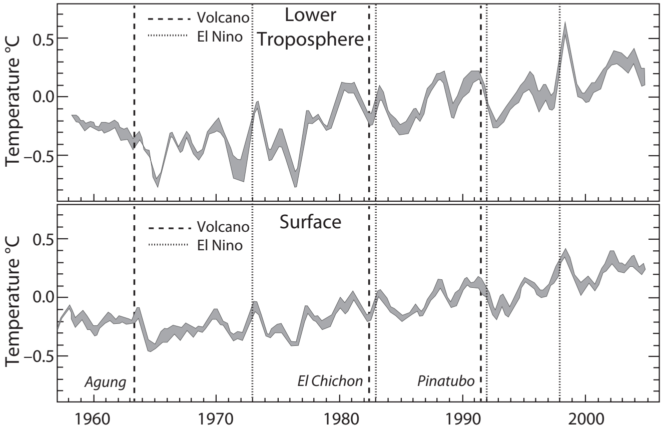


Figure 7.2. Top: Lower troposphere temperature as measured by various satellites and by radiosondes; the gray shading indicates the spread between all measurements. Bottom: Surface temperature records from NOAA, NASA, and UKMO, with gray shading again indicating the spread. Records are monthly means, smoothed with a seven-month running mean filter, and are relative to 1979–1997 mean. Adapted from Solomon et al., 2007.

measurements show trends of between 0.14°C and 0.18°C per decade, depending on the particular method used to obtain temperatures from the brightness measurements.

GLOBAL WARMING IN CONTEXT: THE PAST MILLENNIUM

When discussing global warming, a fair question to ask is, "How does the temperature increase of the past

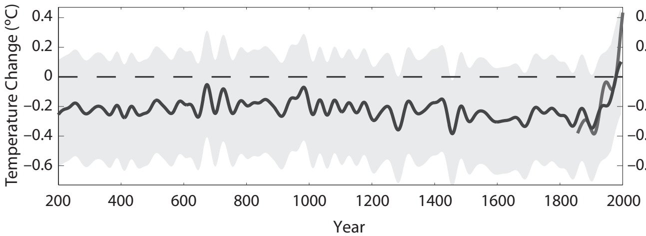


Figure 7.3. Global mean surface temperatures of the past 1,800 years. The lighter solid curve extending from about 1850 to 2000 shows the instrumental record. The longer solid curve is an estimate of temperature over the entire period using proxy reconstructions, and the gray shading is an error estimate (the 95 percent confidence interval). The series are smoothed to remove fluctuations of periods shorter than 40 years, and the temperatures represent anomalies in °C from a late twentieth century value.
Source: Adapted from Jones and Mann (2004).

combined in such a way as to provide a reconstruction of global temperature even though the proxies themselves do not have a uniform global coverage.

Figure 7.3 suggests that the temperatures of the past millennium until about 1900 were relatively uniform compared to the rapid increase in the twentieth century. Note, though, the hint of a "Medieval Warm Period" from about A.D. 900 to A.D. 1200, when temperatures were a little higher than the millennium average, and a Little Ice Age from about A.D. 1400 to A.D. 1800, when temperatures were a little lower. Evidently, though, the rapidity and sustained nature of the warming of the twentieth century was much greater than in any of the previous are burned, they therefore put CO_2 into the atmosphere that has a *lower* abundance of carbon 14 than the CO_2 already in the atmosphere, and it is indeed found that the ratio of carbon 14 in the atmosphere, relative to the other isotopes, is decreasing at about the right rate to be explained by fossil fuel burning.

Although a continuous record of CO_2 levels from direct measurements goes back only about 50 years, we have a good record of carbon dioxide going much further back, mainly from measurements of CO_2 trapped in bubbles in ice cores in Greenland and Antarctica. There are also a number of somewhat isolated measurements at various periods in the past; for example, a series of measurements were made near Paris from 1876 to 1910. The ice cores reveal that CO_2 levels were about 200 ppm at the last glacial maximum some 20,000 years ago, rising over the course of deglaciation to about 260–270 ppm 10,000 years ago, then slowly rising again to about 280 ppm in 1750, just before the Industrial Revolution.[6] The rate of increase since then, to 390 ppm today, is thus far faster than anything else in the past 10,000 years. The level of methane, another greenhouse gas, has also been increasing over the past few decades, although it has leveled off in the past decade.

POSSIBLE CAUSES OF GLOBAL WARMING

The likely culprit

The atmosphere contains greenhouse gases that absorb and re-emit longwave (infrared) radiation, thus warming

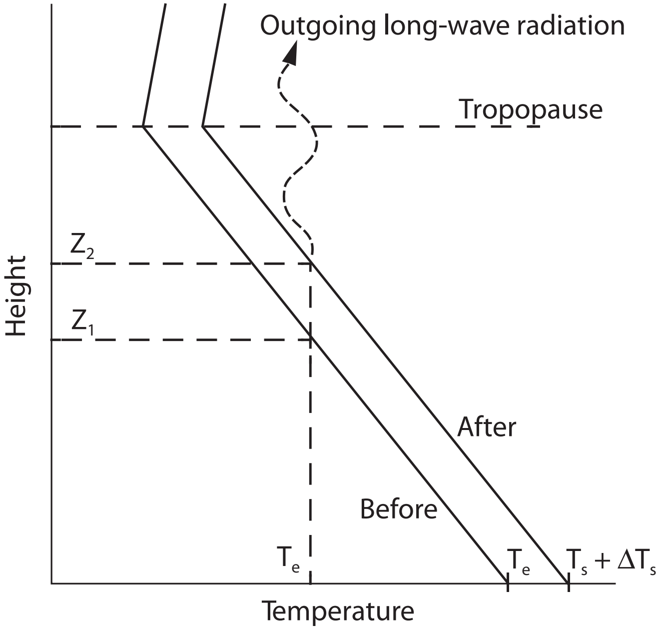


Figure 7.5. Schematic of temperature profiles before and after the addition of greenhouse gases. The total outgoing longwave radiation must remain the same because this radiation balances the incoming solar radiation, and so the emissions temperature, T_e, stays the same. However, the emissions height must increase (from Z_1 to Z_2) because of the increased absorptivity of the atmosphere. Hence, if the temperature gradient in the vertical remains similar, the surface temperature must increase.

levels can have an important effect even if levels of water vapor are already quite high. In this case, it might be thought that since the lower atmosphere is already quite opaque to longwave radiation, the addition of CO_2 would have little effect. However, even in a warm, wet climate because of an increased aerosol concentration in the atmosphere from volcanoes and pollutants, increasing Earth's albedo, possibly in conjunction with natural variability. Regarding the most recent warming, the decade 2000–2009 was 0.2°C warmer than the previous one and about 0.4°C above the 1961–90 average. It was the warmest decade since 1880, and most likely the warmest of the past millennium.[7] The year 1998 was warmed to the tune of about 0.5°C by El Niño, and there have been no comparable El Niño years since, but 2005 and 2010 were nevertheless virtually as warm.

Red herrings and straw men

Of course the models might be wrong and the temperature increase might come from natural causes—natural variability in the climate system, such as we discussed in chapter 6. Thus, the observed warming might not be anthropogenic but rather might be related to our emergence from the Little Ice Age over the past century. This idea is not so much an argument as a speculation because no viable mechanism has been posited that could cause the warming seen over the course of the past century, except possibly for the influence of the ocean, a topic we deal with in the next section. We might imagine that for some reason the ice sheets have retreated, decreasing the albedo and warming the planet, or we might imagine that the clouds have changed configuration in such a way as to cause warming, but there is no evidence for either possibility. If the climate system were so sensitive that such changes were possibilities, we might expect to see evidence of similar changes in past climates. The climate has varied, of course, but looking back at figure 7.3, we see that over the past 1,000 years it has never varied in the same manner as it has in the past 100—the rate of increase of temperature is, so far as we can tell, unprecedented.

One oft-mentioned possibility is that global warming arises because of variations in the sun's output. On the decadal and centennial timescales, variations in the sun's output occur mainly through the solar cycle, a cycle of solar magnetic activity that affects sunspots. The cycle has main periods of about 11 and 22 years, and the former period modulates, albeit slightly, the total solar flux coming into Earth's atmosphere, mainly at ultraviolet wavelengths. The cycle itself cannot cause global warming, but it is possible that there may be longer term variations in the sun's output that modulate the solar cycle, and it is sometimes hypothesized that the Little Ice Age might have been caused by such variations. Indeed, there was a period from about 1645 to 1715, known as the Maunder Minimum, when sunspots seem to have been exceedingly rare, and this period coincided with low temperatures in the middle of the Little Ice Age. Although there is some uncertainty, solar irradiance during the Maunder Minimum is believed to have been about 0.2% less than that of today, or about 0.7 $\mathrm{W\,m^{-2}}$ less.[8] (The change in the ultraviolet component is larger, about 0.7%, but is a small fraction of the total.) The increase in solar irradiance since 1750 is estimated to be at most 0.3 $\mathrm{W\,m^{-2}}$, and the change since 1900, much less than

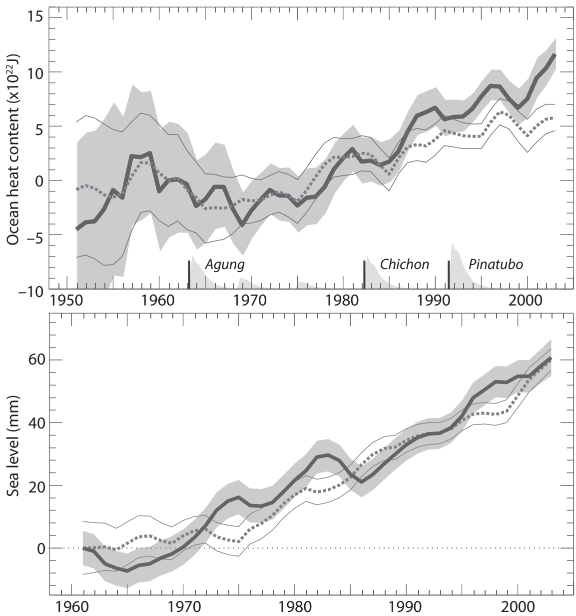


Figure 7.6. Top: The global heat content for the upper 700 m of the ocean (solidblack line, with shading indicating uncertainty) and upper 100 m of the ocean (dotted line, with thin solid lines indicating uncertainty). Bottom: Increase in sea level as estimated from direct measurements (black line and shading) and by combining the contributing components (dotted line and thin solid lines). The time series are all relative to 1961 and smoothed with a three-year running average. Source: Adapted from Domingues et al., 2008.

The total heat capacity of the upper 700 m of the entire ocean is about 10^{24} JK^{-1} (the area of the ocean is approximately 3.6×10^{14} m^2, and the heat capacity of water is about 4×10^3 J kg^{-1} K^{-1}), so that an increase in heat content over the period 1960 to 2000 of 15×10^{22} J

4. The ocean will absorb some of the CO_2 that is added to the atmosphere. he extent to which it does will obviously affect the rate and ultimate level of warming. It will also lead to changes in ocean acidification and possible changes in ocean biology.

Each of these effects is quite important and we devote a section to each, except for the last one, which is beyond our purview. However, we do note that it may take many centuries for the ocean or any other component of the climate system to draw down anthropogenically enhanced levels of CO_2 in the atmosphere.[9]

THE SLOWING OF GLOBAL WARMING

Perhaps the most important and least ambiguous effect of the ocean is that it acts to slow down global warming in a certain sense, although the effect is rather subtle. Furthermore, the ocean is not likely to significantly affect the final equilibrium temperature that the planet will reach if, let us say, CO_2 levels eventually double before leveling off. So just what does the ocean do? Let's try to explain it first just using words; appendix A to this chapter provides a mathematical treatment of the same issue.

In chapter 4 we noted that in the upper ocean there is a mixed layer, typically 50–100 m thick, in which the vertical distribution of temperature and salinity is almost uniform. This turbulent region is stirred by the winds and convection, and its heat capacity is about twenty times that of the atmosphere, which means that it responds rather slowly to such things as daily changes in the weather. However, compared to the deep ocean, it responds very quickly. Thus, for example, if the radiative forcing were to change because of global warming, the mixed layer might be expected to respond and come to a new equilibrium on the timescale of a few years. Now, the radiative forcing has been changing rather slowly over the past century or so, and the mixed layer has little difficulty keeping up. At most, the mixed layer is in equilibrium with the forcing levels in the atmosphere a decade ago, and likely just a few short years ago. The atmosphere, which has a much smaller heat capacity than the mixed layer, tends to respond to the mixed layer quickly. Thus, increased radiative forcing causes the ocean's mixed layer to warm quickly, and this warming in turn sets the atmospheric temperature.

However, the rest of the ocean takes longer—*much* longer—to equilibrate. To get a rough sense of how much longer this time might be, note that in round numbers the mixed layer is 50 m deep, the thermocline is 500 m deep, and the entire ocean, 5,000 m deep; roughly, the times scale accordingly. What effect might these differences have on global warming? Let us suppose that we add some greenhouse gases to the atmosphere and so increase the downward flux of radiation at the surface. Over a few years, the mixed layer warms up until it reaches an equilibrium—that is, a state in which it gives up as much heat as it is receiving—although the temperature of the deep ocean will hardly have risen at all over that period. However, the increased heat that the mixed layer is giving up is going, in part, to warm the ocean below, and it takes a long time, perhaps many centuries, for the deep ocean to fully equilibrate because it is so big. As the deep ocean warms, the mixed layer can give up less of its heat to the ocean below, and so can only balance the radiative forcing by further increasing its temperature, so that it gives its heat back to the atmosphere.

What is the consequence of this picture for global warming? We are slowly but steadily putting greenhouse gases into the atmosphere; the mixed layer responds to this input, and its temperature increases in concert. However, the deep ocean is *far* from equilibrium, which means that, even if we were to stop adding greenhouse gases to the atmosphere and the levels of greenhouse gases were to stabilize at some level, the temperature of the ocean's mixed layer, and so of the atmosphere, would continue to rise for a long period after that. Let us suppose that we continue putting CO_2 into the atmosphere until its level has doubled from that in preindustrial times, and that this doubling occurs in the middle of the twenty-first century. We can expect the global averaged temperature to rise by between 1.3°C and 2.5°C, and probably around 1.8°C, from its preindustrial value by then.[10] Suppose that at that time the political and technological stars align and we are able to prevent greenhouse gas levels in the atmosphere from increasing any further. The average surface temperature of Earth will nevertheless *keep on increasing* until the deep ocean has finally equilibrated, which will take an additional few hundred years or more.

There is considerable uncertainty as to what this final equilibrium temperature rise will be, but it may be much higher than the 1.8°C mentioned above; most estimates are between 2°C and 4.5°C, although higher values cannot be definitively excluded. If we were to eventually cease CO_2 emissions entirely, perhaps because fossil fuels run out, it would take centuries for the CO_2 to finally revert to levels near or a little above the preindustrial value (Archer 2010). During that period, temperatures would likely stay roughly constant for a few centuries before slowly falling back down to levels commensurate with the level of CO_2, as illustrated in figure 7.7. If CO_2 remains doubled for a century before emissions cease, the peak warming will probably be a little over 2°C, and if it triples the peak warming will be around 3°C, with uncertainties of about plus or minus 0.5°C. These peak levels will remain for centuries. Whichever way we slice it, global warming is a long-term problem.

CIRCULATION CHANGES AND A THERMOHALINE SHUTDOWN

One possibility that is raised from time to time is that global warming will bring about a slowdown or cessation of the meridional overturning circulation, sometimes called a *thermohaline shutdown*.[11] Is such a thing likely? First recall that in chapter 4 we divided the ocean circulation up into a quasi-horizontal circulation that is primarily wind driven (the gyres) and an overturning circulation. With warming, the winds may change in

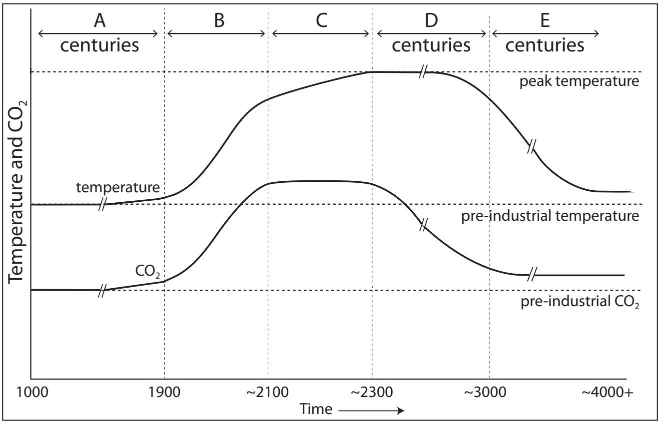


Figure 7.7. Schema of a CO_2-temperature scenario. Carbon dioxide levels increase from 1900 to 2100 (period B) before leveling off (period C) because of controls on emissions. Temperature increases rapidly in period B, then more slowly in period C. At the end of period C (the year 2300 in the figure), anthropogenic emissions go to zero, and the level of CO_2 slowly diminishes through periods D and E back to levels close to, but probably a little above, the preindustrial period. In period D, temperature stays roughly constant for centuries before it too eventually falls back to near pre-industrial levels in period E. Many plausible scenarios can be adapted from this plot by changing 2100 and 2300 to other dates and calibrating the *y*-axis.

detail but almost certainly will not change in their basic structure. Thus, we can confidently predict that the gyres and their western boundary currents will remain qualitatively unchanged in the decades ahead, although certainly there might be shifts in latitude of a few degrees if the surface winds were to change by that amount.

The overturning circulation could, conceivably, change by a larger amount. Recall that one controlling factor in that the deep meridional overturning circulation would significantly weaken and possibly even fall to zero for several decades, and possibly centuries, until the deep ocean temperature warms and the meridional circulation re-establishes itself in the new climate equilibrium. The timescales for the overturning circulation to turn off and back on again are measured in decades and centuries, and if there really were no significant meridional overturning circulation for such a period, the consequences could be severe indeed, with possible wholesale changes in climate at high latitudes. It should be said that most climate models do *not* predict a shutdown in the foreseeable future, but the consequences could be severe if a shutdown were to happen. Global warming makes society confront possibilities that are unlikely but, if they do happen, would produce severe consequences.

SEA-LEVEL RISE

An almost certain consequence of global warming on the ocean is that sea level will rise, if only because as water warms it expands. In the oceans, the only way that an increased volume of the ocean can be accommodated is by an increase in sea level, and sea level has indeed risen over the past several decades, as illustrated in figure 7.6. Sea level is estimated to have risen about 20 cm since records began in the late nineteenth century, and it rose at about 2 mm per year over the last half of the twentieth century, increasing to about 3 mm per year from 1993 to 2003. The fact that sea level has increased over the past century

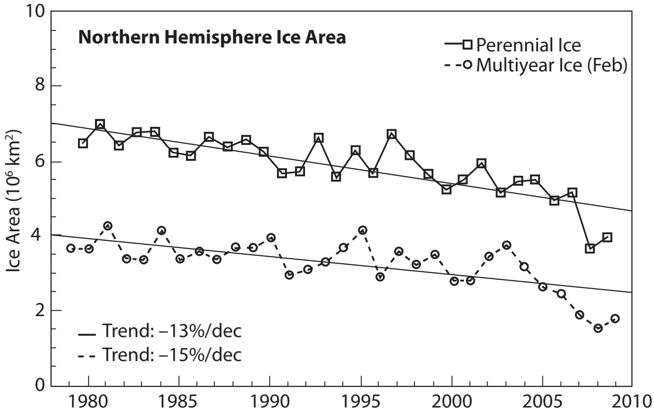


Figure 7.8. Sea ice cover for the Northern Hemisphere from satellite data. Perennial ice excludes seasonal ice cover, and multi-year ice accounts only for ice that has existed for more than one season. Source: Adapted from Comiso, 2002, and Comiso et al., 2008.

climate system were to warm by several degrees Celsius (which would then be quite likely), disappearance cannot be ruled out. Indeed, given that such warming would, as we have discussed, then likely persist for centuries, there is a distinct possibility that the ice sheets on land would also melt, with still more catastrophic consequences.

GLOBAL WARMING—SOME PERSONAL REMARKS

In this last section, I would like to emphasize two aspects about global warming that do not, I think, get sufficient attention: We need to think in terms of probabilities or

GLOBAL WARMING IN A NUTSHELL

Causes and evidence

- Carbon dioxide (CO_2) and a few other gases are greenhouse gases, meaning that they absorb and re-emit longwave radiation that is emitted from Earth's surface, maintaining the surface at a higher temperature than it would be in their absence (about 15°C as opposed to −18°C).
- Greenhouse gases, and in particular CO_2, are added to the atmosphere by the burning of fossil fuels. CO_2 concentration has steadily increased since the beginning of industrialization, from about 270 ppm in 1750 to about 295 ppm in 1900 and 390 ppm in 2010. It will probably exceed 400 ppm some time in 2014.
- The average surface temperature has also increased since preindustrial times and by about 0.8°C over the past century.
- The increase in radiative forcing at the surface is approximately linear—a little less than 4 Wm^{-2} for each doubling of CO_2. This increased forcing (along with other greenhouse gases, and ameliorated by aerosols) can readily account for the observed increase in temperature over the past century.

Projections

- If carbon dioxide levels keep steadily rising and double in the coming century, average temperature increase is expected to be within a range of 1.3°C to 2.5°C, and most likely about 1.8°C, from preindustrial levels by the time of doubling.
- Even if carbon dioxide levels were to stabilize at twice the preindustrial value some time in this century, the temperature would keep on slowly rising to finally reach between about 2°C and 4.5°C higher than the preindustrial value when the ocean equilibrates after a number of centuries.

- If anthropogenic carbon dioxide emissions were completely halted, it would take several centuries for the level of CO_2 in the atmosphere to revert to its preindustrial value. Hence, temperatures would remain high for several centuries.

Other effects on and of the ocean

- Sea level is projected to rise on average by somewhere in the region of 0.4 m over the next century, mostly because of the expansion of seawater as it warms, plus some ice melt, but the uncertainty is large (a factor of 2 either way) and the rise may not be uniform.
- A major melting of the major land-ice sheets is unlikely over the coming century, but if one were to begin, the consequences could eventually be catastrophic, with a sea-level rise of about 6 m if either the Greenland or West Antarctica ice sheets were to completely melt.
- Global warming, once it has occurred, will persist for centuries. Thus, a significant reduction of sea ice and a melting of the Greenland and West Antarctica ice sheets, with concomitant changes in ocean overturning circulation, cannot be ruled out on these timescales. But our ignorance is profound on such matters.

likelihoods, and we need to think clearly about the timescales involved.

Thinking about probabilities is necessary because we don't understand the climate system fully, so we don't know for sure what will happen in the future. The probability we assign to something is then really a measure of the confidence we have in that outcome. However, run over were one in a thousand. The chances might be small, but the consequences, at least to me, would be large. Whether we should live with the risk of potentially catastrophic global warming or "take insurance" by trying to curb emissions today is a question for society as a whole. (Some degree of global warming is of course inevitable.) When dealing with risk, we have to take into account both the likelihood of something happening and the consequences if that something does happen, and we need to weigh the overall risk against the cost of taking insurance. We "buy" insurance by investing in alternative sources of energy, renewable and nuclear, and by living, where possible, less wastefully. There is little downside to this for the developed nations: the cost is not prohibitive compared to the consequences, and in the worst case we prepare for global warming by being efficient and environmentally sound and then the warming turns out to be less than anticipated. For the undeveloped nations and emerging economies such as China and India, the transition to an economy less dependent on fossil fuels may be far more problematic.

The second aspect is the one of timescales, and this is where the ocean in particular comes in. Over the next several decades it is quite plausible (although take heed of the previous paragraphs!) that global warming will continue at about the pace we have seen in the last century. If, let us say, carbon dioxide levels increase in the atmosphere at about 1% per year, then in somewhat less than half a century, they will reach double their preindustrial level and we might expect temperatures carbon dioxide levels may well go up by a factor of six or more with a corresponding warming of almost certainly more than 5°C and possibly more than 10°C, staying at that level for centuries and giving the great ice sheets on Greenland and West Antarctica plenty of time to melt.

Given all the above, one scenario of the future is the following. Let us suppose that we continue to burn fossil fuels for the next few decades, but (rather optimistically) let us also suppose we make good efforts to curb emissions and finally succeed in doing so a few decades hence, and that we are able to stabilize the level of greenhouse gases in the atmosphere at about double the preindustrial level. On this timescale (the short timescale, given the nature of the problem), the additional global warming will likely be a degree or so Celsius, and although there will be some significant regional changes (perhaps especially in precipitation and in extremes of climate), climate change overall may well be less dramatic than the dire scenarios that are sometimes portrayed in the media. However, in the longer term (several decades to centuries and beyond) the problem may be worse that is often expected because global warming will continue relentlessly with consequences to match. An eventual 3°C rise in temperature, which is quite likely if carbon dioxide levels modestly double and stay doubled, will have very large consequences if and when it persists for centuries. Eventually, of course, emissions will diminish or cease if only because fossil fuels run out or become uneconomical, and a scenario for that is illustrated in figure 7.7. If CO_2 levels were to increase to double (or triple) the present value and emissions were, optimistically, to completely cease a century or two after that, the temperature would remain at more than 2°C (or 3°C for tripling) above the preindustrial value for several centuries.

Set against this bleak scenario is the likelihood that society itself will evolve in unforeseen ways, potentially making our current fears moot. Perhaps, then, we simply should not plan for the long term? Perhaps, as Estragon said so memorably in *Waiting for Godot,* there is "Nothing to be done," as the short term will not be so bad and in the long term human development itself is unpredictable. But if we do nothing, then almost certainly carbon dioxide levels will more than double this century and may well double again the following century. Unless we are somehow able to engineer our way out of trouble (for example, by trapping and sequestering the CO_2 emitted when fossil fuels are burned, or extracting CO_2 from the atmosphere), the climate change that will inevitably follow will almost certainly significantly affect the planet Earth itself and all the life on it.

APPENDIX A: MATHEMATICS OF THE TWO-BOX MODEL

Here we give a mathematical description of the two-box model of the ocean, illustrated in figure 7.9. The evolution equations of the two boxes are

$$C_m \frac{\mathrm{d}T_m}{\mathrm{d}t} = F - \lambda_1 T_m - \lambda_2 (T_m - T_d), \tag{7.2a}$$

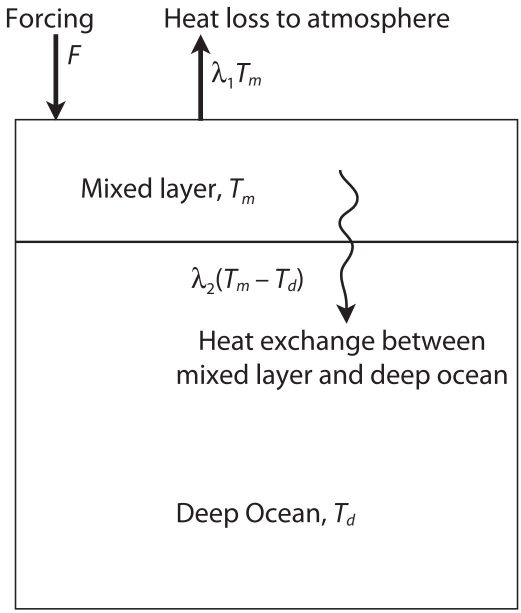


Figure 7.9 A simple two-box model of the ocean, with a mixed layer at a temperature T_m and a deep ocean layer at a temperature T_d, and exchanges of heat between the components as shown.

$$C_d \frac{\mathrm{d}T_d}{\mathrm{d}t} = \lambda_2(T_m - T_d). \tag{7.2b}$$

In these equations, t is time, T_m and T_d are the temperature anomalies of the mixed layer and deep ocean, respectively, F is the anomalous radiative forcing caused by greenhouse warming, and C_m and C_d are the heat capacities of the mixed layer and deep ocean, respectively. The parameters λ_1 and λ_2 are exchange coefficients that determine the rate at which heat is transferred from the upper ocean to the atmosphere and from the upper ocean to the deep ocean, respectively. Although an exact solution of the above equations is often possible (depending on the form of F), it is more informative, and more general, to look at approximate solutions, and that is how we will proceed. The main assumption we make is that the heat capacity of the deep ocean is far greater than that of the mixed layer ($C_d \gg C_m$), which is a good assumption considering that the depth of the mixed layer is typically $\leq$ 100 m, whereas the depth of the ocean itself is on average about 4,000 m.

Given the big disparity in heat capacities, there will be two timescales to the problem: a short timescale over which the mixed layer comes into a quasi-equilibrium, and a much longer timescale over which the full ocean equilibrates. In the short timescale, the deep ocean does not respond and its temperature stays at the initial temperature, namely zero (because all temperatures are measured relative to the initial temperature). Equation 7.2a becomes

$$C_m \frac{\mathrm{d}T_m}{\mathrm{d}t} = F - (\lambda_1 + \lambda_2)\, T_m. \tag{7.3}$$

If we turn on the forcing and then hold it constant, the solution of equation 7.3 is found to be

$$T_m = \frac{F}{\lambda^*}(1 - \mathrm{e}^{-t\lambda^*/C_m}), \tag{7.4}$$

where $\lambda^* = \lambda_1 + \lambda_2$. There are two conclusions to be drawn at this stage:

1. The system evolves toward a quasi-equilibrium on a short timescale of $t_s = C_m/\lambda^*$. Observations and experiments with comprehensive climate models suggest that this timescale is on the order of a few years to a decade.
2. The quasi-equilibrium temperature reached on this short timescale is given by $T_2 = 0$ and

$$T_m = \frac{F}{\lambda_1 + \lambda_2}. \tag{7.5}$$

Let us now consider timescales much longer than C_m/λ^*. We suppose that the mixed layer is in a quasi-equilibrium and that equation 7.2 may be approximated by

$$0 = F - \lambda_1 T_m - \lambda_2 (T_m - T_d), \tag{7.6a}$$

$$C_d \frac{\mathrm{d}T_d}{\mathrm{d}t} = \lambda_2 (T_m - T_d). \tag{7.6b}$$

The mixed-layer temperature is thus given by

$$T_m = \frac{\lambda_2 T_2 + F}{\lambda_1 + \lambda_2}, \tag{7.7}$$

and substituting this in equation 7.6b gives

$$C_d \frac{\mathrm{d}T_d}{\mathrm{d}t} = \frac{\lambda_2}{\lambda^*} (F - \lambda_1 T_2). \tag{7.8}$$

The system now evolves on the long timescale $t_l = C_d \lambda^* / (\lambda_1 \lambda_2)$, which, given the large value of C_d, may be measured in centuries. The final equilibrium reached has the temperature

$$T_m = T_d = \frac{F}{\lambda_1}, \tag{7.9}$$

which is higher than the temperature given by equation 7.5.

Let us end with few cautionary notes and general remarks. First, and to summarize, there is a fast evolution to the temperature $F/(\lambda_1 + \lambda_2)$, followed by a much slower evolution to the final temperature F/λ_1. However, the real ocean does not consist of just two boxes; the real ocean is immensely complicated. There is indeed a big separation between the timescale on which the mixed layer responds and that on which the full ocean equilibrates, but there are a number of intermediate timescales on which other aspects of the oceans respond, like the gyre and the thermocline. So our treatment is a simplification. Nevertheless, it does contain an essential truth, and that is that it will take a long time for the ocean to equilibrate fully. Even if we were to curtail our emissions of greenhouse gases into the atmosphere and the levels were to stop increasing, the temperature would slowly keep on increasing, possibly for hundreds of years, until the true equilibrium were reached.

Further Reading

MAINLY THE ATMOSPHERE

Wallace, J. M. & Hobbes, P., 2006. *Atmospheric Science: An Introductory Survey*. 2d ed. Burlington, Mass., Academic Press. Covers a wide range of topics at the advanced undergraduate level.

Andrews, D. G., 2010. *An Introduction to Atmospheric Physics.* 2d ed. Cambridge, U.K., Cambridge Univ. Press. Written at the upper-division undergraduate level, with a couple of chapters on dynamics.

Holton, J. R., 2004. *An Introduction to Dynamic Meteorology.* 4th ed. Burlington, Mass., Academic Press. A textbook on dynamics at the undergraduate and graduate levels.

MAINLY THE OCEAN

Denny, M., 2008. *How the Ocean Works: An Introduction to Oceanography.* Princeton, N.J., Princeton Univ. Press. Discusses the ocean from a mechanistic point of view, including physical, chemical, and biological aspects.

An Open University Course Team, 1998. *The Ocean Basins: Their Structure and Evolution.* 2d ed. Oxford, U.K., Pergamon Press.

An Open University Course Team, 2001. *Ocean Circulation.* 2d ed. Oxford, U.K., Pergamon Press. The Open University has a series of books on various aspects of earth sciences written at the undergraduate level.

Pickard, G. L. & Emery, W. J., 1988. *Descriptive Physical Oceanography: An Introduction.* 5th ed. Oxford, U.K., Butterworth-Heinemann. A descriptive survey of the oceans, ocean basins, seawater, and circulation.

Knauss, J. A., 1997. *Introduction to Physical Oceanography.* 2d ed. Long Grove, Ill., Waveland Press. Covers some of the same ground as Pickard and Emery, but with more emphasis on the underlying physical principles and dynamics.

CLIMATE

Bigg, G., 2003. *The Oceans and Climate.* 2d ed. Cambridge, U.K., Cambridge Univ. Press. Goes beyond this book by covering the chemical and biological, as well as physical, interactions.

Kump, L. R., Kasting, J. F. & Crane, R. G., 2009. *The Earth System.* 3d ed. Prentice Hall. A book on the Earth system as a whole, from paleoclimate and ecosystems to ocean circulation and global warming.

Hartmann, D. L., 1994. *Global Physical Climatology*. San Diego, Calif., Academic Press. Covers the physical principles of the climate system, from the fundamental principles to how the global climate system works as a whole.

Marshall, J. C. & Plumb, R. A., 2008. *Atmosphere, Ocean, and Climate Dynamics: An Introductory Text*. Burlington, Mass., Elsevier Academic Press. Covers a variety of topics in climate dynamics at a level appropriate for advanced undergraduates or beginning graduate students.

http://www.realclimate.org/ This website has many informative, and sometimes pointed, blogs on climate issues. As with any website, the links may not last indefinitely.

Glossary

This glossary gives an informal description of some of the main technical terms used in the book.

Abyss: The deep ocean, extending from the thermocline to the seafloor. Sometimes, the abyss is taken to be only the deepest part of that region, with the middepth ocean extending from immediately beneath the thermocline to about 2 km deep and with the abyss beneath this.

Aerosols: Particulate matter in the atmosphere with both anthropogenic origins (e.g., pollution) and natural origins (e.g., volcanoes).

Baroclinic instability: A hydrodynamic instability that gives rise to weather in the atmosphere and to mesoscale eddies in the ocean.

Carbon dioxide, CO_2: A trace gas, consisting of two oxygen atoms bonded to one carbon atom, comprising about 0.039% of the atmosphere. It is a major greenhouse gas, second only to water vapor in its effect in the current atmosphere, but it is the single most important cause of the greenhouse effect.

Centrifugal force: An apparent force that acts on all bodies in a rotating frame of reference. The force acts to