see that a *zonal* wind stress (i.e., a nonzero τ_w^x) induces a *meridional* flow in the ocean, and a meridional wind stress induces a zonal flow. That is, the average induced velocity in the Ekman layer—the *Ekman transport*—is perpendicular to the imposed wind stress at the surface. In the Northern Hemisphere where f is positive, if the stress is eastward (i.e., τ_w^x is positive), then the Ekman transport is southward (V is negative).

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

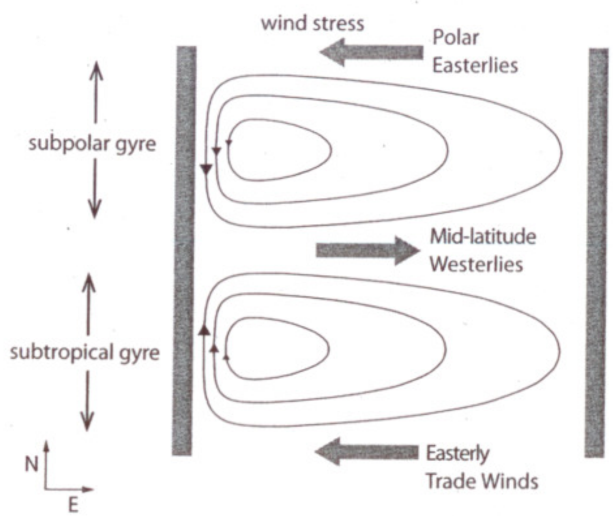


Figure 4.1. An idealized gyre circulation in a rectangular ocean basin in the Northern Hemisphere, showing the subtropical gyre (lower, typically extending from about 15°N to 45°N), the subpolar gyre (upper), and the intense western boundary currents on the left.

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

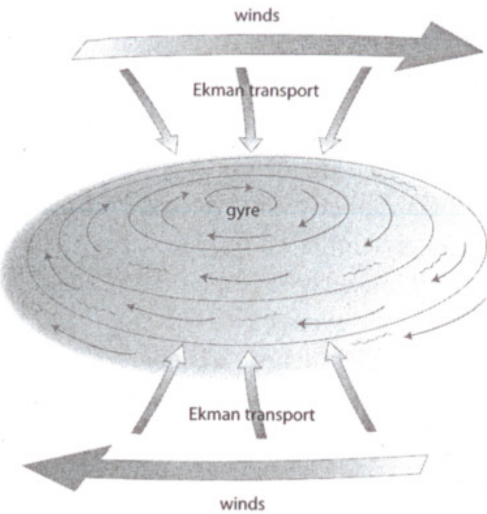


Figure 4.2. Production of gyres by winds. The winds blowing as shown induce a converging Ekman flow, causing the sea level to increase in the center, thus giving rise to a pressure gradient. This gradient in turn induces a geostrophic flow around the gyre, in the same sense as the winds themselves.

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

are strongest on the equatorial side. Now the Ekman transport is directed *away* from the center of the gyre, and the sea level is depressed and upwelling occurs.

The doming of the sea surface produces a pressure gradient in the ocean, as illustrated in figure 4.2. Consider a horizontal plane at a level a little below the sea surface. The pressure at that level is produced by the weight of the fluid above it, as we discovered in the section in chapter 3 on hydrostatic balance, and so is higher where the sea surface is higher. This pressure gradient produces a geostrophic flow perpendicular to the pressure gradient, and so in the same direction as the wind that originally produced the doming. Thus, when all is said and done, on a rotating planet the wind leads to the production of an ocean current that is aligned with the wind, rather as we would expect in the nonrotating case. However, the pressure gradients in the two cases are quite different because of the presence of the Coriolis force in the rotating case; note in particular that the horizontal pressure gradient produced by the doming extends all the way to the bottom of the ocean. Thus, even though the direct effects of the wind stress are confined to the upper few tens of meters, the wind produces geostrophic currents that can extend to great depths.

Sea-surface slope and the geostrophic current

It may seem a little fantastical that the wind can produce a change in the sea level and that this in turn can produce the currents of the great gyres. However, we do not need a large change in the sea level to produce quite substantial flows, as we can see with a simple calculation. The geostrophic current is a balance between the Coriolis and pressure gradient forces, so that

$$fu = -\frac{1}{\rho}\frac{\partial p}{\partial y}, \qquad -fv = -\frac{1}{\rho}\frac{\partial p}{\partial x}. \tag{4.1a, b}$$

The pressure at a level below the surface is given by the weight of the fluid above it, so that

$$p = \rho g h, \tag{4.2}$$

where h is the height of the column of seawater and ρ is the density of the seawater. Thus, using equation 4.2 in equation 4.1a and b, we get

$$fu = -g\frac{\partial h}{\partial y}, \qquad fv = g\frac{\partial h}{\partial x}. \tag{4.3a, b}$$

Suppose that the height of the sea surface varies by just 1 m over a horizontal distance of 1,000 km—a truly small slope that would be extremely difficult to detect by measurements of the ocean surface but that is, remarkably, measurable using modern satellites. The magnitude of the currents produced is then given by

$$u = \frac{g}{f}\frac{\Delta h}{L} = \frac{9.8}{10^{-4}}\frac{1}{10^{6}} \approx 0.1\ \mathrm{m\,s^{-1}}. \tag{4.4}$$

Such a current is easily measurable, and when one considers that billions of tons of water might be put in motion this way, one begins to see the large effect that this current can have.

WESTERN INTENSIFICATION

We have now explained the underlying reason for gyres, but we have not explained one of their most important and indeed obvious aspects: the gyres are not symmetric in the east–west direction. Thus far, our explanation would lead to gyres that look like those in the left panel of figure 4.3, whereas in fact the gyres look more like those in the right-hand panel, with intense *western boundary currents,* of which the Gulf Stream in the western North Atlantic is the most famous example to Americans and Europeans, and the Kuroshio is the corresponding current off the coast of Japan. The presence of the Gulf Stream has been known for a long time—Benjamin Franklin was one of the first people to chart it. So our question is a simple one: why is the Gulf Stream in the west?

It turns out that the cause of the western intensification is that, as we discussed in chapter 3, the magnitude of the effective rate of rotation (specifically, the magnitude

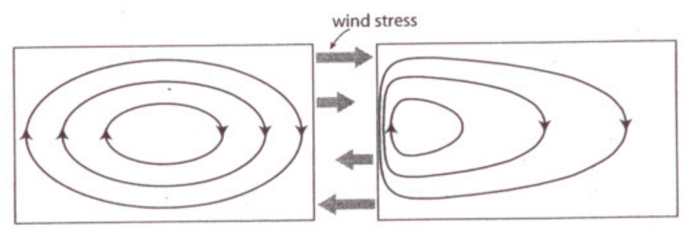


Figure 4.3. Two schematics of a subtropical gyre. The left panel shows the basic response of the circulation to the winds shown, and the right panel shows the gyres in the presence of differential rotation, with western intensification.

of the Coriolis parameter) of Earth increases as we go poleward. This *differential rotation* causes the gyres to have a marked east–west asymmetry, with the flow in the west squished up against the coast. As the effect is both important and hard to grasp, we give a couple of explications. For definiteness we focus on the subtropical gyre in the Northern Hemisphere, but the same principles apply to the other gyres.

Torques and interior flow

If the wind stress acting on the ocean varies with latitude—as we see that it does in figure 4.3—then the wind provides a torque that tends to *spin* the ocean. In a steady state, not only do the forces on the ocean have to balance but so do the torques; otherwise the ocean would spin faster and faster. The torques on the ocean are provided by the wind, by friction, and by the Coriolis force (the pressure gradient does not provide a torque).[2] Integrated over the entire ocean basin, the wind torque is balanced by the frictional torque, and since the frictional torque normally acts in a sense opposite to that of the spin of the fluid itself, the basin-scale circulation spins in the same sense as the wind. That is, for there to be a balance between wind and friction, the large-scale flow must have the same overall sense of rotation as the wind, producing the gyre shown in the left panel of figure 4.3.

However, in the interior of the basin, frictional effects are in fact very weak and the spin provided by the wind stress is locally balanced by the effects of the Coriolis

force. Now, the Coriolis parameter increases northward, and it turns out that to locally balance the wind torque, a meridional flow must be produced in the ocean interior. The direction of the meridional flow depends on the sense of the spin provided by the wind, but in the subtropical gyre the meridional flow turns out to be equatorward. Let's see why.

Consider a parcel of fluid in the middle of the ocean, as illustrated in figure 4.4. The wind blows zonally with a stronger eastward wind to the north and so provides a clockwise torque. We can balance this torque by a Coriolis force if there is a *southward* flow of water in the interior. In that case, the Coriolis force provides a westward force on all the parcels of fluid moving south. However, the force is stronger on the fluid that is in the northern part of the domain (because the Coriolis parameter increases northward), so the spin provided to the fluid is counterclockwise, opposing that spin provided by the wind. The southward flow adjusts itself so that the spin provided by the varying Coriolis force just balances the spin provided by the wind. (The balance is called *Sverdrup balance*, and the southward flow is called the Sverdrup interior.)

Now, the southward-flowing water must return northward somewhere, and this return must be at either the eastern or western boundaries because here the frictional effects of the water rubbing against the continental shelf and coast are potentially able to allow the flow to achieve a torque balance and move northward. However, only if the boundary layer is in the west (as illustrated in the right panel of figure 4.3) can such a balance be achieved: the

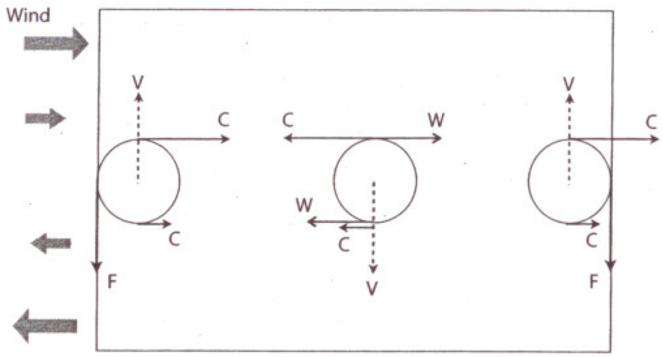


Figure 4.4. The production of a western boundary current. Schematic of the torques (namely, the spin-inducing forces: the wind, W; Coriolis, C; and friction, F) acting on parcels of water in the ocean interior (center) and western and eastern boundary layers (left and right), in a Northern Hemisphere subtropical gyre. In the interior, friction is small and the torques balance if the flow (denoted V) is southward. If the northward return flow is in the west, then a balance can be achieved between friction and Coriolis forces, as shown. If the northward return flow is in the east, no balance can be achieved.

gyres then circulate in the same sense as the wind forcing, and the frictional forces at the western boundary act to oppose the wind forcing and achieve an overall balance. If the return flow were to be in the east, then the flow would, perversely, be circulating in the opposite sense to the torque provided by the wind, and no balance could be achieved. A local view of how the torque balances work in the boundary layer is provided in figure 4.4.

Suppose that the wind blew the opposite way. The balance of the wind torque and the Coriolis effect can now be achieved if interior flow is northward, and this is the case

in the subpolar gyres. For the overall flow to have the same sense as the wind torque, the return flow still has to be in the west. Thus, we see that western boundary currents are a consequence of the differential rotation of Earth, not the way the wind blows. If Earth rotated in the opposite direction, the boundary currents would be in the east.

Westward drift

In this section, we give a slightly different explication of why the boundary current is in the west. It is not really a different explanation because the cause is still differential rotation, but here we think about it quite differently. We'll see that the effect of differential rotation is to make patterns propagate to the west, and hence the response to the wind's forcing piles us in the west and produces a boundary current there.

We noted already that the component of Earth's rotation in the local vertical direction also increases as we move northward or, putting it a little informally, the *spin* increases northward. (The spin is also called the *vorticity*.) Now consider a parcel of fluid sitting in the ocean. It may be spinning from two causes, namely, because it is spinning relative to Earth and because Earth itself is spinning. If that parcel moves and if no external forces act upon it, then the total spin of the fluid parcel is preserved. Its local spin relative to Earth must therefore change to compensate for changes in Earth's spin.

Let's now imagine a line of parcels, as illustrated in figure 4.5. Suppose we displace parcel A northward.

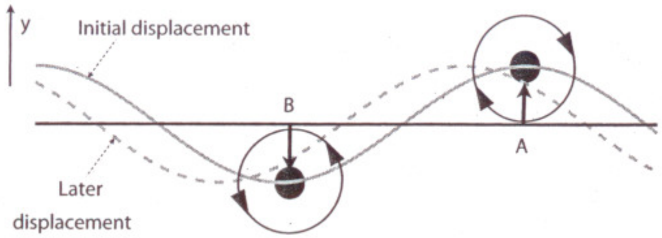


Figure 4.5. If parcel A is displaced northward, then its clockwise spin increases, causing the northward displacement of parcels that are to the west of A. A similar phenomenon occurs if parcel B is displaced south. Thus, the initial pattern of displacement propagates westward.

Because Earth's spin is counterclockwise (looking down on the North Pole) and this spin increases as the parcel moves northward, then the parcel must spin more in a clockwise direction to preserve its total spin. This spin has the effect of moving the fluid that is just to the west of the original parcel northward, and then this fluid spins more clockwise, moving the fluid to its left northward, and so on. The northward displacement thus propagates westward, whereas parcels to the east of the original displacement are returned to their original position so that there is no systematic propagation to the east. Similarly, a parcel that is displaced southward (parcel B) also causes the pattern to move westward. This is an idealized example—in fact we have just described the westward propagation of a simple *Rossby wave*—but the same effect occurs with more complex patterns and in particular, with the gyre as a whole. Thus, imagine that an east–west symmetric gyre is set up, as in the left panel

of figure 4.3, with the winds and friction in equilibrium. Differential rotation then tries to move the pattern westward, but of course the entire pattern cannot move to the west because there is a coastline in the way! The gyre thus squishes up against the western boundary in the manner illustrated in the right panel figure 4.3, creating an intense western boundary current. This way of viewing the matter serves to emphasize that it is not the frictional effects that cause western intensification; rather, frictional effects allow the flow to come into equilibrium with an intense western boundary current, with the ultimate cause being the westward propagation caused by differential rotation.

THE OVERTURNING CIRCULATION

The other main component of the ocean circulation is the *meridional overturning circulation* (MOC), circulation essentially occurring in the meridional plane. There are two rather distinct aspects to this circulation, but they each have a common feature, namely the sinking of dense water at high latitudes and its subsequent rise to the surface elsewhere. Thus, in general the overturning circulation may be regarded as being "buoyancy enabled" in the sense that without buoyancy gradients at the surface there would be no deep overturning circulation. The buoyancy gradients themselves are produced by variations in temperature and salinity, and so the circulation is also sometimes known as the thermohaline circulation. The two different aspects are the processes that keep the water circulating. In one case, it is mixing by small-scale turbulent motions, and in the other case, it is the direct effect of the wind. We'll deal with these in turn, but before that, we discuss the buoyancy force itself.

The buoyancy or Archimedean force

The force due to buoyancy is one of the most familiar forces occurring in a fluid and, rather famously, was known to Archimedes. It is the force that, among other things, allows objects to flow in water. The Archimedes principle is often stated as "Any object, partially or wholly immersed in a fluid, experiences an upward force equal to the weight of the fluid displaced by the object." Let's see why this is so.

Consider a container of still water and focus attention on a particular piece of water that is fully surrounded by other fluid. The parcel has a finite weight, of course, and it does not sink to the bottom of the container because it is held up by the pressure force provided by the rest of the fluid in the container. Because none of the water is moving, the weight of the parcel (its mass times the acceleration due to gravity, acting downward) must exactly equal the upward pressure forces provided by the rest of the fluid. Now, let us replace the parcel with a solid object of the same shape and size. The upward pressure force provided by the rest of the fluid remains the same; this, we just ascertained, is equal to the weight of the parcel of fluid displaced—and this is Archimedes' principle. If the solid object is lighter than the weight of the fluid

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.

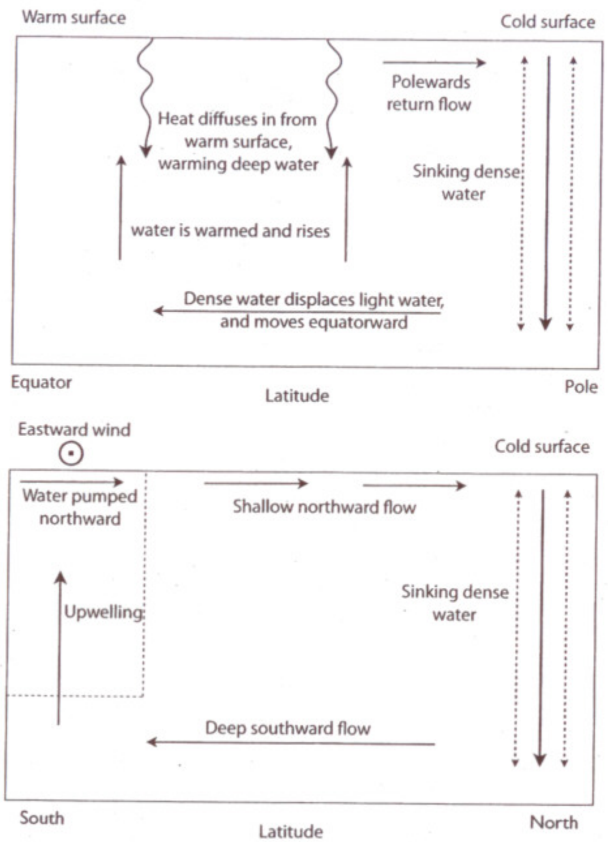


Figure 4.6. Schema of the two main components of the MOC. Top: The mixing-maintained circulation. Dense water at high latitudes sinks and moves equatorward, displacing warmer, lighter water. The cold, deep water is slowly warmed by diffusive heat transfer (mixing) from the surface in mid- and low latitudes, 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 *left* 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

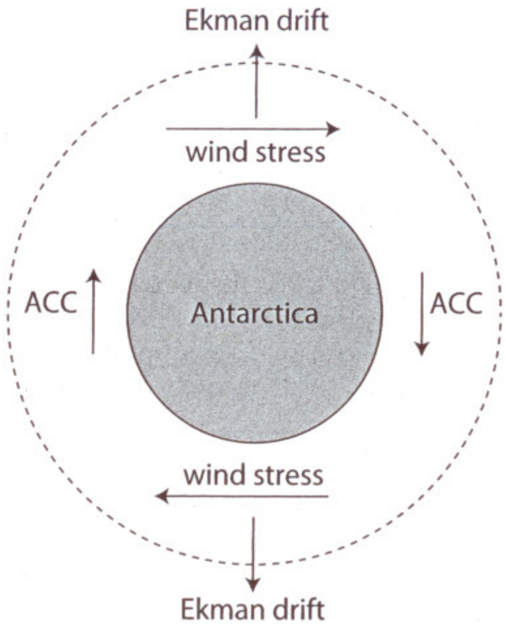


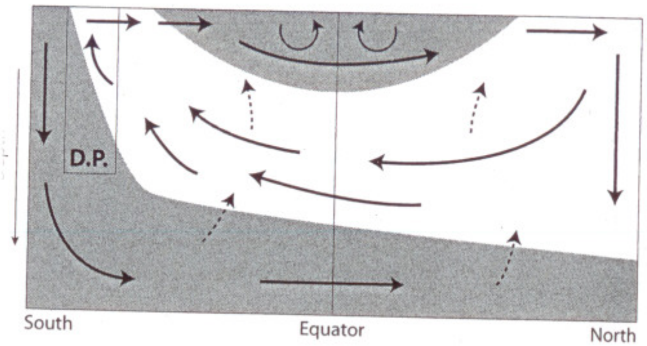
Figure 4.7. Schematic of the flow in the Antarctic Circumpolar Current (ACC). The wind predominantly blows in a zonal direction around the Antarctic continent, generating an Ekman flow toward the north and a net loss of water from the channel. The water returns at depth, generating a deep overturning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



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

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

OCEAN CIRCULATION IN A NUTSHELL

he large-scale ocean circulation may usefully be divided into quasi-horizontal circulation, comprising the gyres and other urface and near-surface currents, and a meridional overturn- ng 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

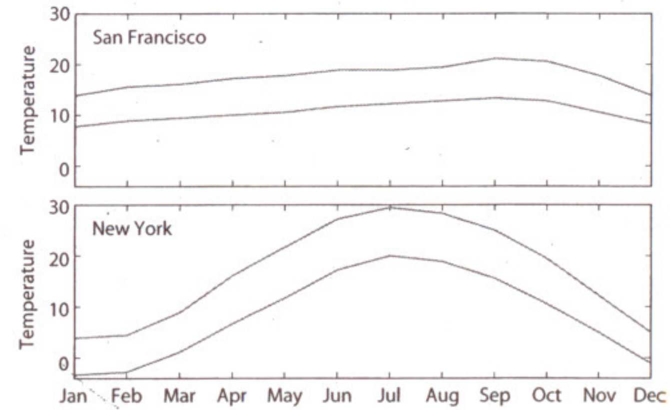


Figure 5.1. The seasonal cycle of temperature (°C) in San Francisco and New York. For each city, we plot the average low temperature and the average high temperature for each month. Note the much bigger range in New York and the maximum earlier in the year, in July rather than September.

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 constant motion, and so the depth of ocean being heated and cooled over the seasonal cycle is much larger than the depth of land that is being heated and cooled. Let's explain that in a bit more detail.

First of all, the higher the heat capacity of a body, the more heat is needed to change its temperature. Thus, if an object is being cyclically heated and cooled, as in a seasonal cycle, then the change in its temperature is much smaller if its heat capacity is higher. Now, to what depth in the land and ocean does the heat penetrate over the course of a seasonal cycle? Plainly not all of the ocean's great depth is heated during the day or cooled during the night, or even over the course of a season. Rather, regarding the ocean, just that part that is turbulently mixed by the effects of wind (and in part by heating and cooling itself) fully partakes in the annual cycle, namely the *mixed layer,* which we discussed in chapter 2. Although its character varies from place to place in the ocean, it has a typical depth of about 50–100 m. That is, the effective heat capacity of the ocean is approximately that of a body of water 50–100 m deep. This heat capacity is quite large, and for comparison, the heat capacity of the atmosphere corresponds to a depth of just 3 m of water.

What is the effective heat capacity of land? Two effects make it much less than that of water. First, the specific heat of dry land is about 4 times less than that of water (for wet land, the factor is about 2). Second, because land is, rather obviously, not in motion in the same way as the

ocean is, the penetration of heat into the land is much less than it is into the ocean. The heat can penetrate only by conduction, and because the earth (soil) has rather low thermal conductivity, only the top few meters are significantly heated and cooled over the course of a season. The same can be said for the major ice sheets over land, such as those over Greenland and Antarctica: they have large mass but low thermal conductivity. Thus, combining the effects of a larger heat capacity and a larger effective depth, the ocean has an effective heat capacity that is about 100 times greater than that of land. This high heat capacity considerably attenuates the seasonal cycle and is a good part of the reason for the large difference between San Francisco and New York.

San Francisco is a rather extreme case because not only is the summer temperature moderated by the ocean, but also the interminable fog that blows in from the ocean and covers the city like a wet blanket keeps the summer temperatures miserably low and makes them seem even lower, as Mark Twain perhaps felt. However, the heat capacity effect does occur on very large scales. The surface of the Southern Hemisphere is about 80% ocean whereas the Northern Hemisphere is only about 60% ocean, and as a consequence the seasonal cycle is much more pronounced over the Northern Hemisphere than the Southern, as we see in figure 5.2. Between 40° N and 60° N, the amplitude of the annual cycle is about 12°C, whereas between the corresponding latitudes in the Southern Hemisphere, the annual cycle has an amplitude of only 3°C.

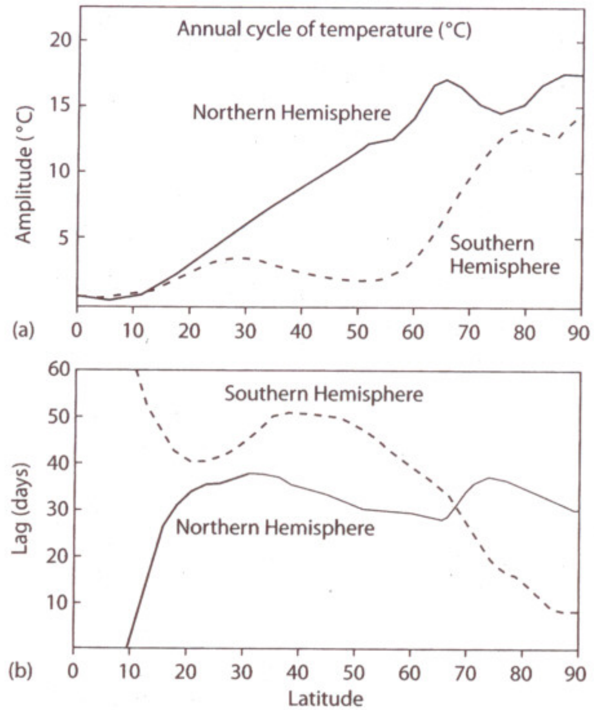


Figure 5.2. Amplitude and lag of the annual cycle in the Northern and Southern hemispheres, as a function of latitude. The lag is the time, in days, from the maximum solar insolation to the maximum temperature. Source: Trenberth, 1983.

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

$$C\frac{dT}{dt} = S - \lambda T. \qquad (5.1)$$

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

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

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 transport in Earth's atmosphere may be calculated directly because we are constantly taking measurements of the air temperature and its flow for weather forecasts. One way to then determine the total heat transport by the atmosphere at a given latitude is to sum up the product of the temperature and meridional velocity over all longitudes and over the entire depth of the atmosphere. Given the heat transport by both the atmosphere and by the atmosphere–ocean system, the heat transport by the ocean follows by simple subtraction, and this transport is shown in the dashed line in figure 5.3a. It is also possible to calculate the ocean transport directly, using in situ ocean measurements; the advantage is that one may be able to elucidate the individual mechanisms of ocean heat transport rather than just the overall effect. Such direct measurements tend to be less accurate than the residual method because of the sparsity of measurements in the ocean, but the two methods are broadly consistent.

Let's first look at the total heat transport. Evidently in mid- and high latitudes the atmospheric transport is two to three times that of the ocean, whereas in low latitudes the two are comparable, with the ocean exceeding that of the atmosphere at very low latitudes. The atmospheric heat transport, which we won't consider in any detail, takes place via two main mechanisms: In low latitudes, the transport occurs via the zonally symmetric Hadley cell, which takes warm air poleward and cooler air equatorward. In midlatitudes, the heat transport in the atmosphere occurs through the familiar weather systems,

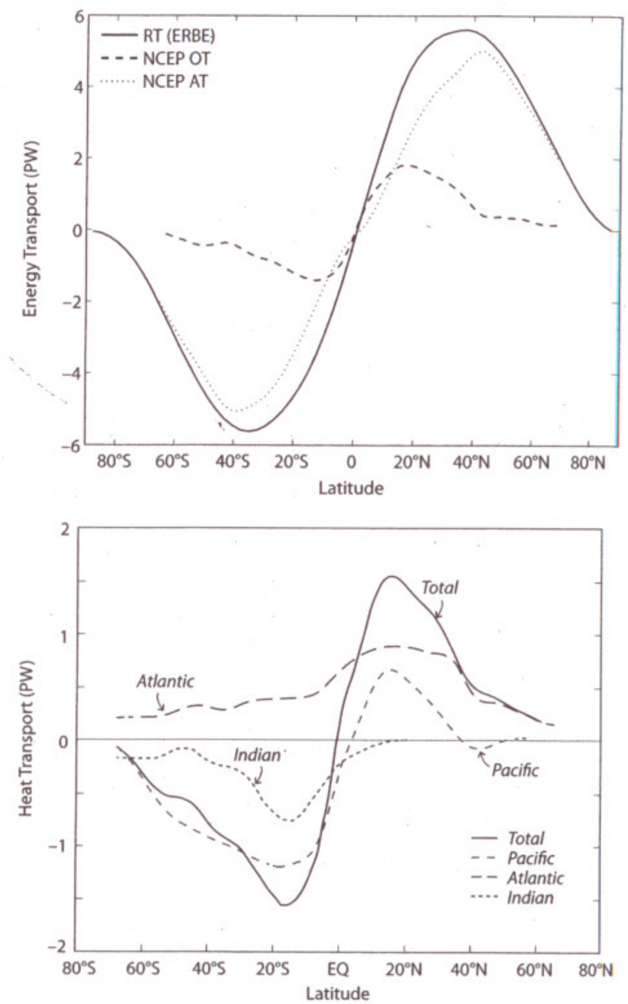


Figure 5.3. Upper panel: Heat transport in the total atmosphere–ocean system (solid line), in the ocean (dashed line), and in the atmosphere (dotted line). Lower panel: Oceanic heat transport, subdivided into the various basins. Source: Trenberth and Caron, 2004.

which are continually stirring the atmosphere and bringing warm air with low-latitude origins poleward and cool, high-latitude air equatorward.

As for the ocean overall, and as we might expect, it transports heat poleward in both hemispheres. The transport is associated with a release of heat into the atmosphere at high latitudes, whereas the ocean is being heated by the atmosphere at low latitudes. Measurements show that the release of heat occurs in two primary locations: at the poleward end of western boundary currents, notably in the western Atlantic and western Pacific oceans at about 40° north and south, and at very high latitudes, in particular in the North Atlantic around Greenland. Another notable feature about the ocean transport is that the poleward transport is much larger in the Northern Hemisphere than in the Southern. In fact, if we look at the contributions from the individual basins in the right-hand panel of figure 5.3, we see that the heat transport is northward (that is, *toward* the equator) in the South Atlantic! Such a transport is quite remarkable, for it implies that the ocean is not being thermally driven by the meridional temperature gradient alone (which would transport heat from hot places to cold places). What could be the driving force? If we recall our discussion of the oceanic meridional circulation in the previous chapter, we won't be too surprised to discover that it is the wind, but not the wind-driven gyres. Let's look at the mechanisms of ocean heat transport in a bit more detail and try to sort this out.

What mechanisms?

Oceanic heat transport is mainly effected by the large-scale circulation, with some transport by mesoscale eddies, mainly in the ACC. As we discussed in chapter 4, there are two distinct aspects to this circulation: the wind-driven gyres and upper ocean circulation, and the meridional overturning circulation. Let's see how each of these transport heat.

The wind-driven gyres

The wind-driven gyres, especially the subtropical wind-driven gyres, are a major factor in the heat transport in both hemispheres. If we consider the North Atlantic as an example, poleward heat transport occurs because the western boundary current (the Gulf Stream in this case) brings warm water up from the tropics along the eastern seaboard of the United States, releasing heat (especially in winter) when the warm water comes into contact with the colder air coming off the cold continental land mass. Such a release of heat occurs at the western edges of all the major ocean basins in midlatitudes, for example, off the coasts of Japan, the Eastern United States, South America (south Brazil, Uruguay, and Argentina) and southeastern Australia. The poleward flow of the western boundary currents is balanced by equatorial flow in the middle of the gyres that brings cold water equatorward, although the flow is broader and weaker so that the transport of cool water equatorward is spread over a large area. But, in any case, the consequence of warm water flowing poleward in the western boundary currents and cool water flowing equatorward in the interior means that the wind-driven gyres transport heat poleward. The transport occurs in both the Pacific and the Atlantic, and in both the Northern and Southern hemispheres. The subpolar gyres in the Northern Hemisphere also transport poleward, but they are less well defined than the subtropical gyres and cover less of the ocean so that their heat transport is somewhat weaker than that of the subtropical gyres.

The meridional overturning circulation

We saw in chapter 4 that the meridional overturning circulation has two mechanistically distinct components: a mixing-maintained component and a wind-maintained component, and the net overturning circulation is a combination of the two. The mixing-maintained component of overturning circulation responds to the buoyancy gradient at the surface between the equator and the pole, and in today's climate that buoyancy gradient is mainly a consequence of the temperature gradient. As described in chapter 4, the buoyancy gradient leads to a circulation in which cold water at high latitudes sinks and moves equatorward, balanced by warmer, near-surface water moving poleward. The net effect of this circulation is a poleward transport of heat that would be, in the absence of other effects, roughly equal in magnitude in the two hemispheres.

The other component of the overturning circulation is a pole-to-pole circulation, driven by the wind in the southern oceans. In the Atlantic Ocean, this component tends to dominate the purely buoyancy-driven circulation, as suggested by figure 4.8 in chapter 4. We see from this figure that there is a large northward, and so poleward, heat transport in the Northern Hemisphere because the equatorward moving water has come from high northern latitudes and is correspondingly cold, and the poleward moving water nearer the surface is warmer. In the Southern Hemisphere, however, the southward moving water is colder than the northward moving water because the cold North Atlantic Deep Water continues its path into the Southern Ocean and the water moving northward near the surface is relatively warm. That is, the water moving equatorward is generally warmer than the water moving poleward! Thus, the heat transport in the Atlantic from the deep circulation is northward in both hemispheres. The wind-driven gyres, of course, transport heat poleward in both hemispheres, partly counteracting the equatorward heat transport by the deep circulation in the Southern Hemisphere, so that the net effect is that the heat transport is weak and equatorward in the South Atlantic, whereas it is strong and poleward in the North Atlantic. In the Pacific and Indian ocean basins, there is no corresponding wind-driven deep circulation that transports heat northward, and the wind-driven and buoyancy-driven circulations both act to transport heat poleward, as we can see in figure 5.3.

What are the effects?

What are the gross effects of the ocean heat transport on the climate? The main effect is simply that the high latitudes, especially the high latitudes in the Northern Hemisphere, are warmer than they would be if the oceans were not present. How much warmer is a question that we cannot answer with armchair reasoning. We would need to perform detailed calculations with comprehensive climate models of the type used to predict the weather or used for global warming experiments.

One such set of experiments was performed by M. Winton of the Geophysical Fluid Dynamics Laboratory in Princeton, and we briefly describe some of the results found (Winton 2003). Climate models solve the equations that determine the temperature and motion of both the atmosphere and the ocean. The models also have representations of sea ice and cloudiness and of their effects on the incoming solar radiation and outgoing infrared radiation. Thus, for example, snow and ice cover reflects solar radiation back to space, making the climate cooler than it would be in their absence. If the ice sheets were for some reason to expand, the climate would cool, the ice sheets would further expand, and the climate would further cool—an example of a positive feedback, in this case the *ice–albedo feedback*.

In one set of numerical experiments, the ocean was replaced by a mixed layer, so that although the heat uptake and release of the ocean throughout the annual cycle are

accounted for, the effect of the ocean currents and so the oceanic heat transport are completely removed. When the ocean heat transport is removed, the atmosphere tries to compensate for this change by transporting more heat poleward itself. However, in spite of this and in spite of the relatively small heat transport of the oceans compared to the atmosphere at high latitudes, the effects of the oceans are found to be quite large because of a feedback involving sea ice and, to a lesser extent, low-level cloudiness. The simulations without oceanic heat transport all developed large ice sheets that covered mid- and high latitudes, making the overall climate much colder than it is now.

What seems to happen is the following. Although the atmosphere is able to partially compensate for the lack of an ocean transport, the atmospheric transport naturally occurs at a higher elevation than the ocean transport. The lack of an ocean heat transport enables sea ice to grow, and once the ice begins to grow, the positive ice–albedo feedback comes into play and the ice grows more. The detailed mechanism for the strong effect of the ocean seems to involve the upward convective flux of heat in the wintertime: The meridional overturning circulation leads to convection at high latitudes, with cold water parcels sinking and being replaced by slightly warmer parcels, which then release heat into the atmosphere. This process is eliminated when the ocean is replaced by a mixed layer, allowing sea ice to grow.

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

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

The amplitude of the cycle is $S_0/C\omega$, which is small because C is large. That is, because $C\omega > \lambda$ in this case, the variations in temperature are smaller than they are in the case with slow forcing variations given by equation 5.6. Because high frequencies are damped more than low frequencies, we say that the variations are *reddened*. Note too that the temperature variations are now out of phase with the heating. Thus, the maximum temperature occurs when the heating is least. It is this effect that accounts for the delay in the maximum temperature in maritime climates, with the hottest part of the year occurring in late summer or even the beginning of autumn.