Princeton Primers in Climate

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CLIMATE AND THE OCEANS

Geoffrey K. Vallis

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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

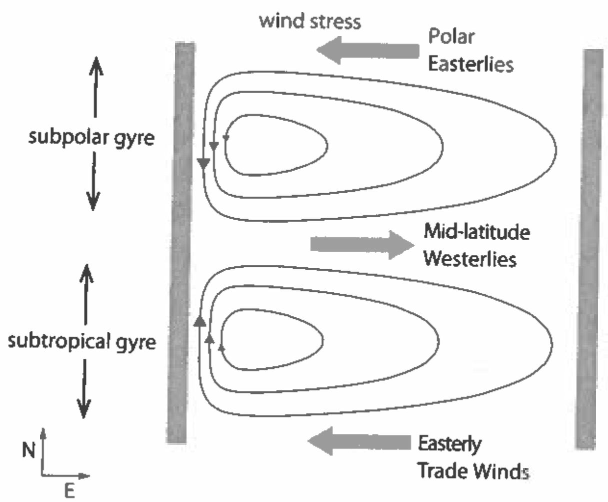

Figure 4.1. An idealized gyre circulation in a rectangular ocean basin in the Northern Hemisphere, showing the subtropical gyre (lower, typically extending from about 15°N to 45°N), the subpolar gyre (upper), and the intense western boundary currents on the left.

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

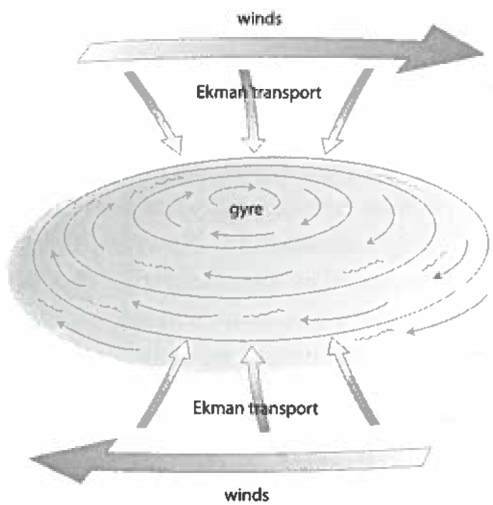

Figure 4.2. Production of gyres by winds. The winds blowing as shown induce a converging Ekman flow, causing the sea level to increase in the center, thus giving rise to a pressure gradient. This gradient in turn induces a geostrophic flow around the gyre, in the same sense as the winds themselves.

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

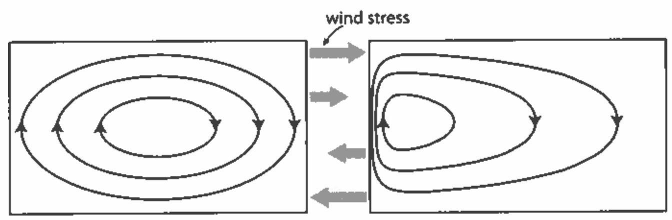

Figure 4.3. Two schematics of a subtropical gyre. The left panel shows the basic response of the circulation to the winds shown, and the right panel shows the gyres in the presence of differential rotation, with western intensification.

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

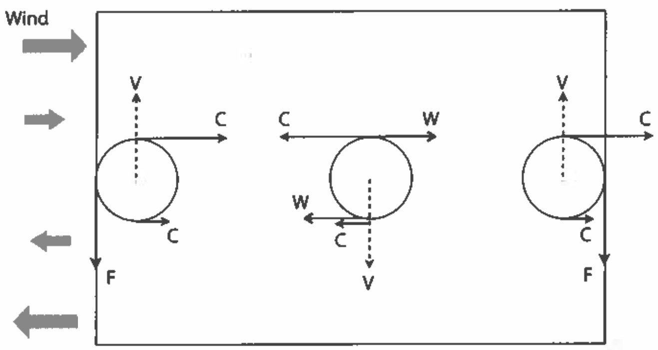

Figure 4.4. The production of a western boundary current. Schematic of the torques (namely, the spin-inducing forces: the wind, W; Coriolis, C; and friction, F) acting on parcels of water in the ocean interior (center) and western and eastern boundary layers (left and right), in a Northern Hemisphere 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.

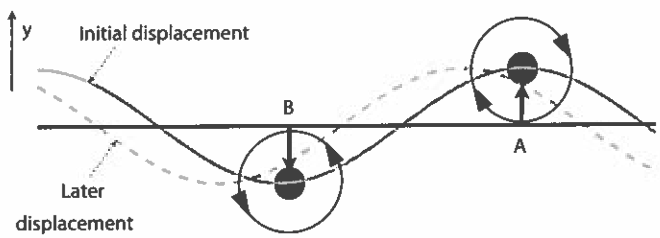

Figure 4.5. If parcel A is displaced northward, then its clockwise spin increases, causing the northward displacement of parcels that are to the west of A. A similar phenomenon occurs if parcel B is displaced south. Thus, the initial pattern of displacement propagates westward.

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.

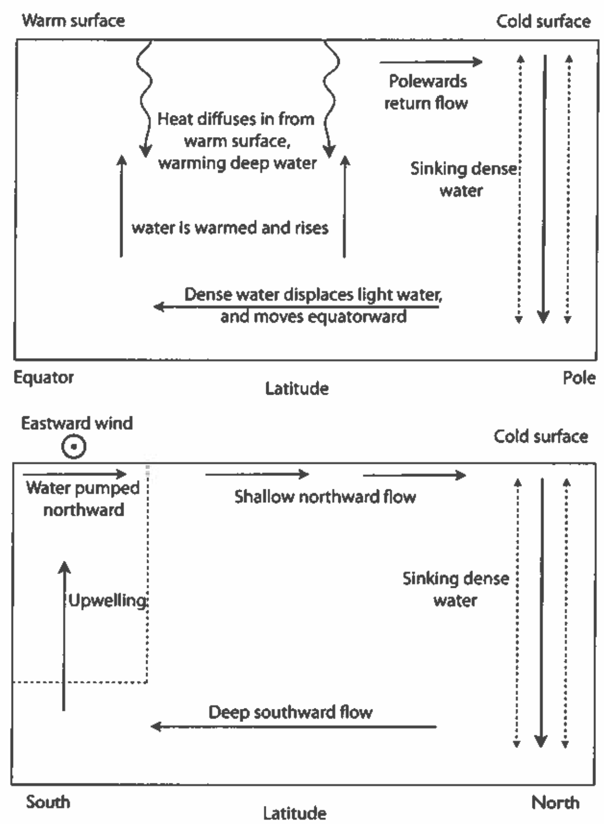

Figure 4.6. Schema of the two main components of the MOC. Top: The mixing-maintained circulation. Dense water at high latitudes sinks and moves equatorward, displacing warmer, lighter water. The cold, deep water is slowly warmed by diffusive heat transfer (mixing) from the surface in mid- and low latitudes,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

what about Coriolis?

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

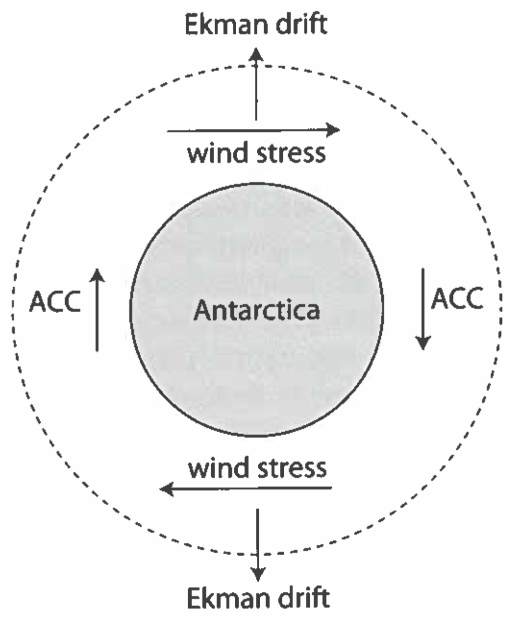

Figure 4.7. Schematic of the flow in the Antarctic Circumpolar Current (ACC). The wind predominantly blows in a zonal direction around the Antarctic continent, generating an Ekman flow toward the north and a net loss of water from the channel. The water returns at depth, generating a deep overturning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

diabatic route vs. adiabatic route

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

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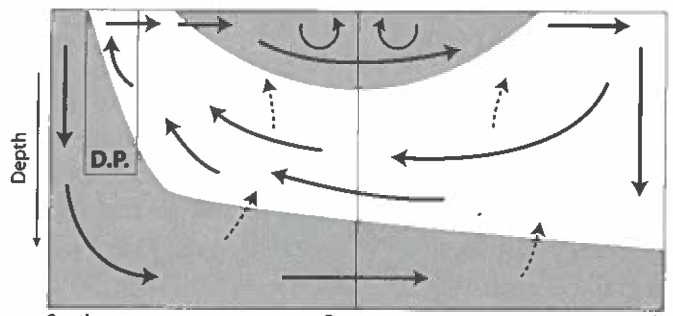

Figure 4.8. Schematic of the meridional overturning circulation, most applicable to the Atlantic Ocean (D.P. indicates the Drake Passage, the narrowest part of the ACC). The arrows indicate water flow, and dashed lines signify water crossing constant-density surfaces, made possible by mixing. The upper 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

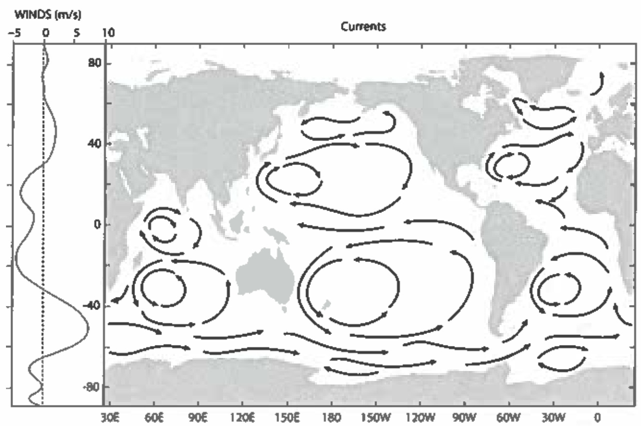

Figure 2.3. A schematic of the main surface currents of the world's oceans. The panel at the left shows the zonally averaged zonal (i.e., east–west) surface winds.

the circulation is dominated by the *subtropical gyres*. The flow in the subtropical gyres is westward on their equatorial branch and eastward in midlatitudes on their poleward branch, and it is not too difficult to imagine that this circulation is directly driven by the wind—the westerlies (i.e., wind from the west) in midlatitudes and the generally easterly (wind from the east) trade winds in low latitudes. There are subtropical gyres in both the Atlantic and Pacific oceans in both hemispheres (figure 2.3), as well as in the southern Indian Ocean. The circulation in these gyres occurs mainly in the upper few hundred meters of the ocean, weakening and eventually becoming unrecognizable in the deep abyss.

A rather conspicuous aspect of the subtropical gyres is their east–west asymmetry: the meridional component of all of these gyres is much stronger in the west of the oceans, giving rise to what are called *western boundary currents*. In the North Atlantic, this current is called the Gulf Stream, in the North Pacific, the Kuroshio, in the South Atlantic, the Brazil Current, in the South Pacific, the East Australia Current, and in the Indian Ocean, the Agulhas Current. The cores of these currents are often only 100 km or so wide, and the speed of the flow can reach a quite tidy 1 m/s, which is not at all sluggish for the ocean, where most currents are more like a few centimeters per second. The equatorward return flow in all the subtropical gyres is spread over a much greater longitudinal extent and so is much weaker.

In the Northern Hemisphere, poleward of the subtropical gyres, the circulation consists of *subpolar gyres*. Because of the converging meridians and the complicated geography of the North Atlantic and North Pacific, these gyres are not nearly as conspicuous as their subtropical counterparts, but they too are primarily wind driven, a consequence of the strong midlatitude westerly winds and the weak easterly winds at high latitudes. They also have intense western boundary currents, now flowing equatorward, and it may be useful to look ahead to see a schematic of how the ocean gyres might be if the oceans were purely rectangular in figure 4.1 in chapter 4. The reader may imagine how the circulation would be distorted, but might keep its essential structure, if the exist in all the major basins, and the Atlantic has a robust overturning circulation. How does it all fit together? There is no universally accepted picture, and certainly no quantitative theory. Rather, observations and numerical simulations have been used to put together a qualitative picture, as illustrated in figure 2.6. The circulation illustrated should be regarded as a highly schematic representation, perhaps even as a metaphor, of the real ocean circulation, in part because of the presence of ocean eddies discussed below. Also, only the more global features are illustrated; thus, we see a cross-hemispheric circulation in the Atlantic, but not its vertical structure, which we'll talk about more in chapter 4. However, the figure does illustrate the main features of the global circulation: a meridional overturning circulation, the sinking of cold dense water in the North Atlantic and off Antarctica, and the western boundary currents.

Ocean eddies

We are all familiar with the fact that the weather differs from the climate, the difference arising because the atmospheric flow is unsteady, and we talk more about this in chapter 6. The same applies to the ocean, only more so: the large-scale currents in the ocean are almost all unstable, rather like a river flowing over rapids, and tend to break up into smaller *mesoscale eddies*, as illustrated in the lower panel of figure 2.6. The resulting eddies are the oceanic analogue of atmospheric weather, although because of differences in the physical properties of the

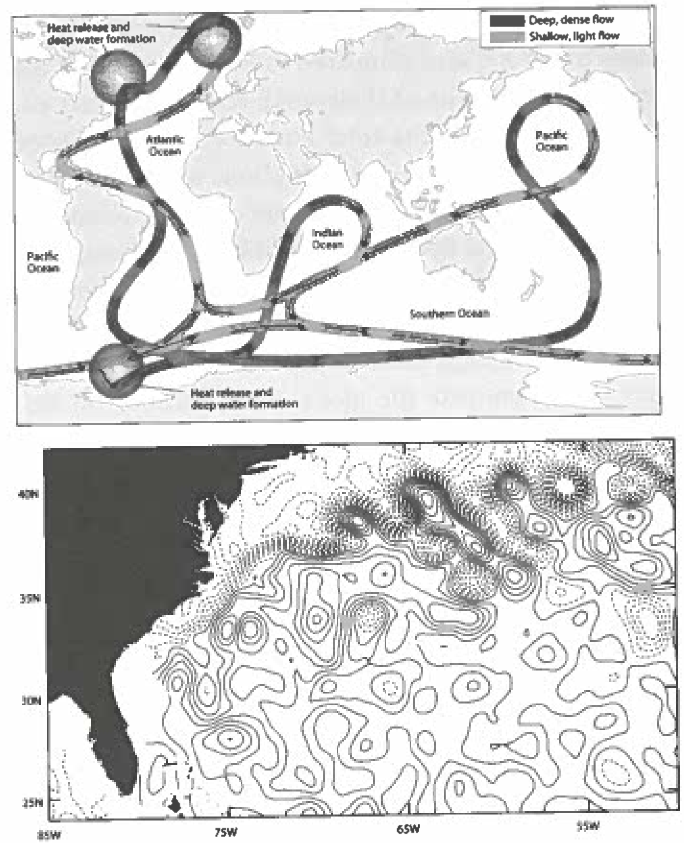

Figure 2.6. Top: An artist's impression of the global ocean circulation, sometimes called the "conveyor belt." Bottom: The sea-surface height in the Atlantic on October 15, 2008, indicating the presence of the Gulf Stream and mesoscale eddies.