Physics of the Ocean and Climate

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Introduction

Today we'll talk about some of the physics of the ocean and some of the ways that it affects the climate. One of the reasons that the ocean is interesting is because it is *vast*, not only physically, but conceptually. Its phenomena cover a huge range of spatial and temporal scales, all of which interact, and it integrates all types of science: physics, chemistry, biology, and geology, all of which contribute to many processes. In this talk, we will mostly focus on the physics of the large scales. Two of the central concerns of physical oceanography are large-scale ocean circulation and turbulence. We will focus on the former, and I'll give a separate talk on turbulence next year.

1 Overview of the Ocean

Geography

From the perspective of the ocean, the world map makes the most sense centered on Antarctica. The "center" of the ocean is the Southern Ocean surrounding the Antarctic continent, which supports the largest current in the world, and the only one that can propagate unimpeded all the way around a circle of latitude. Extending outward from the Southern Ocean are three basins: the Pacific, Atlantic, and Indian, from largest to smallest. The North Atlantic connects to the Arctic Ocean, which can be thought of as a small ocean or as an "estuary" of the Atlantic. (The other connection from the Arctic to the world ocean is the Bering Strait, which is much too shallow to allow for meaningful currents to connect it to the Pacific.)

Atmosphere and Forcing

It is impossible to properly discuss the ocean without the atmosphere. The two fluid bodies should be thought of as an inherently coupled system that exchanges momentum, mass, and heat between its two components at the sea surface. The energy input that drives the system is solar energy. Of the incoming radiation, about 30% is reflected, while 20% is absorbed by the atmosphere and the remaining 50% by the ocean. The ocean is a fluid primarily heated from above, leading to a stable stratification of light warm water over cold dense water. But the atmosphere is a fluid heated primarily from below, making it constantly unstable to convection.

The ocean therefore generates its own conditions for motion by transferring a large portion of its heat to the convecting atmosphere, where the resulting winds drive the surface ocean circulation. The majority of this heat transfer is accomplished via the latent heat of evaporating water, particularly in the tropics. We will talk about some details of this wind-driven circulation in a few minutes.

The heating of the ocean also directly generates currents by setting up horizontal density gradients. Because the tropics receive more radiation than the polar regions, the water there is warmer and lighter than the polar water. This creates a pressure gradient, which on a non-rotating planet would create a stable equatorward surface current with a compensating poleward deep current. Things are more complicated on Earth since it rotates. We'll talk about that in a bit.

Scales

The ocean covers a huge range of spatial and temporal scales, each with its own physical phenomena. The nonlinearity of the Navier-Stokes equation allows different scales to talk to each other, so that energy imparted by wind stress at larger scales eventually makes its way to be dissipated at the smallest scales. We will be focusing on scales from about 100 km to 10,000 km, which comprise the "large-scale" circulation of the ocean. But it should be emphasized that these phenomena are inherently coupled to those at smaller scales through the turbulent energy cascade.

Vertical Structure

I should also mention the vertical structure of the ocean - there is a warm layer at the surface on the order of 100m deep. This is called the 'mixed layer', as it is continually mixed by wind stress at the surface. Between about 100m and 1000m are the thermocline and pycnocline, where the ocean rapidly becomes colder and denser. Below about 1000m is the abyss, where the properties of the ocean are relatively constant down to 4000m depth.

We call the density variation with depth "stratification", and it imparts a large amount of structure to ocean circulation since it takes energy to cross a density gradient. Water generally moves along isopycnals, surfaces of constant density, and only crosses them via turbulent mixing. If not for turbulence, anomalies would only travel from the surface ocean to the deep ocean via molecular diffusion, which takes up to a million years.

An additional consideration is the disparity between the horizontal and vertical scales. The small aspect ratio of the oceans means that in most flows, the vertical and horizontal components of the flow behave differently and it is warranted to treat them separately.

2 Large-Scale Dynamics

Transport

The physics of the ocean is fundamentally concerned with transport of quantities throughout the ocean. Again, there are a huge number of quantities that we could track, but for the purposes of physics, there are only a few quantities that we are primarily concerned with:

- momentum
- mass
- temperature
- salinity

The phenomena that contribute to transport are generally advection and diffusion. Transport equations take the form of conservation equations that balance these transport phenomena with the rate of change of a quantity and its sources and sinks.

The momentum conservation equation, or Navier-Stokes equation, is just Newton's second law for a fluid. For a Newtonian fluid, it looks like

$$\frac{\partial \vec{v}}{\partial t} + (\vec{v} \cdot \nabla)\,\vec{v} = -\frac{1}{\rho}\nabla p + \nu \nabla^2 \vec{v} + \vec{F}$$

The second term on the LHS represents the self-advection of momentum, and it is the source of nonlinear and turbulent behavior in fluids. In practice, we split the velocity into a macroscopic part and a microscopic part, and the microscopic advection becomes an additional momentum diffusion term on the RHS, combining with the viscosity term. In general, this turbulent diffusion dominates mixing.

Along with the momentum equation is the mass conservation equation, known as the continuity equation

$$\frac{\partial \rho}{\partial t} + \nabla \cdot (\rho \vec{v}) = 0$$

Often we can take our fluid to be incompressible, meaning that the density is a constant. In this case, the continuity equation becomes

$$\nabla \cdot \vec{v} = 0$$

This is four equations in five unknowns, so more information is required. We introduce an equation of state

$$\rho = \rho\,(P, T, S)$$

where S is the salinity. The equation of state for seawater (or water, for that matter) doesn't have a closed-form expression, like the ideal gas law, but it can often be sufficiently approximated by linear deviations from a reference state.

We have now introduced temperature and salinity into our system, meaning that we need more equations. For the salinity, we may add an advection-diffusion equation to account for its dynamics. For temperature, we must introduce a thermodynamic relation. This is more complicated for seawater than it is for an ideal gas, so we will not address it now.

These are the fundamental equations for transport of conserved quantities in seawater. Of course, the familiar conserved quantity that's missing here is angular momentum. It will appear in a different form a bit later.

Geostrophy

Now we'll come to the actual motion of the ocean. What makes it so interesting and unintuitive is the fact that the rotation of the Earth gives rise to a Coriolis term in the momentum equation that varies with latitude. This additional term looks like

$$F_{\text{cor}} = 2\vec{\Omega} \times \vec{v}$$

At small scales, the effects of rotation aren't visible, but when we look at large scales, fluids behave in surprising ways.

When studying fluid flow, we generally work locally on a tangent plane centered at some latitude θ_0. We may approximate the coriolis term by a Taylor series. The "f-plane" approximation takes only the constant term and says that

$$|F_{\text{cor}}| = 2\Omega \sin\theta_0 v = f_0 v$$

while the "β-plane" approximation includes the first order variation:

$$|F_{\text{cor}}| = (f_0 + \beta y)\, v$$

We'll examine the momentum equation in the f-plane approximation at large Rossby number, meaning that the scale is large enough that the effects of rotation dominate the advection term in the equations of motion. The Rossby number is just the ratio of the sizes of the advection term and coriolis term:

$$\text{Ro} = \frac{|(\vec{v} \cdot \nabla)\, \vec{v}|}{\left|2\vec{\Omega} \times \vec{v}\right|} \sim \frac{U}{fL}$$

So for large enough length scales, we expect rotation to dominate.

We'll treat our fluid as inviscid and work with only the horizontal motions, so that gravity isn't relevant. What does a steady state flow look like in this case? The horizontal momentum equation takes the form

$$f\hat{z} \times \vec{u} = -\frac{1}{\rho}\nabla_z p$$

This describes rotational flow where the sea surface slopes in a way to create a pressure gradient - either a hill or a depression, and the current flows along isobars around the center. The slope of these hills is extremely small, on the order of 1 micron per meter. But this is enough to balance the coriolis force.

We can also take the curl of the above equation, which after a bit of simplification gives

$$(\hat{z} \cdot \nabla)\, \vec{u} = 0$$

This says that the velocity is constant in the vertical direction. This goes by the name of the Taylor-Proudman effect, and it says that rotating fluids exhibit much more structure than non-rotating ones.

The situation gets more complicated if we allow for stratification, meaning that we allow the density to vary with depth. In this case, we end up with the 'thermal wind' relation, where there is an additional balance between vertical gradients of horizontal velocity and horizontal gradients of density.

Another feature we can add is to upgrade to the β-plane. This is where consideration of angular momentum becomes very relevant. In the case of fluids, we work instead with vorticity, which for horizontal flow looks like

$$f + |\nabla \times \vec{v}|$$

a sum of a 'planetary vorticity' and a 'relative vorticity'. On a β-plane, poleward movement increases planetary vorticity, which must be balanced in the equation of motion for vorticity. We won't write that one, but instead the equation of motion for 'potential vorticity', which is much nicer:

$$\frac{D}{Dt}\left(\frac{f + |\nabla \times \vec{v}|}{H}\right) = 0$$

I.e. potential vorticity is conserved (assuming no friction or heating).

One effect this has on the flow is 'western intensification': the flow becomes stronger at the western edge of the basins, i.e. off the east coast of continents. This is the phenomenon where western boundary currents such as the Gulf stream off the US east coast and the Kuroshio off Japan are much stronger than their corresponding eastern boundary currents, such as the California current or Canary current.

Ekman Flow

Now we'll get to the question of how a wind stress creates currents. In this case, it is the wind stress and eddy viscosity instead of the pressure gradient that balances the coriolis force. Our horizontal equation is

$$f\hat{z} \times \vec{u} = \frac{1}{\rho}\frac{\partial \vec{\tau}}{\partial z}$$

For the wind stress, $|\tau|$ is proportional to the square of the wind speed. Below the surface, it is instead the eddy viscosity that balances the coriolis force, and this is parameterized as

$$\vec{\tau} = \rho A_z \frac{\partial \vec{u}}{\partial z}$$

Thus we get the equation

$$f\hat{z} \times \vec{u} = A_z \frac{\partial^2 \vec{u}}{\partial z^2}$$

The solution to this equation is the Ekman spiral, where each layer of fluid pulls on the one above it and twists it to the right. If we integrate the velocity over the whole column, the net water transport is 90 degrees to the right of the original wind stress.

The wind stress controls vertical motion in the ocean as well through upwelling and downwelling. The continuity equation tells us that regions of positive horizontal divergence will suck up water from the deep ocean, while negative divergence pumps water into the deep ocean. The end result is that the rate of upwelling or downwelling is proportional to the curl of the wind stress:

$$w_E \propto \nabla \times \vec{\tau}$$

Meridional Overturning

We said that wind stress is only one of two processes that generate large-scale ocean circulation. The other is the "thermohaline circulation", where cold, salty, dense water sinks near the poles in a process called "deepwater formation". The primary regions of deepwater formation are in the North Atlantic near Greenland and in the Southern Ocean. The subducting water takes its own characteristics with it, forming distinct "water masses" that can be tracked through the ocean. The water follows deep currents and surface currents, making a full transit of the ocean in about 1000 years.

3 Climate

Greenhouse Effect

Now let's take a few minutes to sketch some connections between the oceans and climate. Again, there are many scales to climate. In short, the global temperature is controlled by the movement of carbon between reservoirs such as the crust, ocean, vegetation, soils, and atmosphere. The reason why carbon is so important is because of the greenhouse effect, which we'll briefly sketch.

Consider the Earth-Sun system without accounting for the atmosphere. If we treat both the Earth and Sun as blackbodies, we can ask the question: what is the equilibrium temperature of Earth? To find this, we'd solve the equation

$$(1-\alpha)\, S_0 \pi R^2 = 4\pi R^2 \sigma T_e^4$$

The answer we find is $T_e = -23$°C. This can't be right! The observed average surface temperature is $T_{\text{obs}} = 15$°C. What we are missing is the greenhouse effect

- not all of Earth's emitted radiation actually makes it to space. Gases in the atmosphere absorb the outgoing radiation and reemit it isotropically, meaning that some is returned to the surface. To escape to space, the radiation has to be continually absorbed and reemitted upwards. But for this to actually warm the planet, the gases must not do the same thing to the incoming radiation from the sun. So a greenhouse gas is one that is strongly absorbing around Earth's outgoing wavelength (i.e. in the IR, or longwave), but transparent around the Sun's incoming wavelength (i.e. in the visible, or shortwave).

The most potent greenhouse gas is water vapor, which strongly absorbs IR radiation. However, we're lucky in that the concentration of water vapor is naturally limited by the fact that it will eventually rain out when the air is saturated. So the most important greenhouse gas becomes the next in line: CO2. This molecule has a vibrational mode near the peak of the Earth's emission spectrum. So the Earth is habitable because of water vapor and CO2 in the atmosphere.

CO2 Storage

The ocean plays important roles in climate both via heat flux and CO2 flux, but we will focus on the latter. The deep ocean is a huge carbon reservoir, and it can take up much of the excess CO2 in the atmosphere. There are several mechanisms for transporting carbon to the deep ocean, and they are dominated by biological processes. About 25% of our emitted carbon each year is taken up by the surface ocean. The timescale for deep ocean circulation is a few hundred to a thousand years, and we can expect that the ocean will take up about 75% of our CO2 emissions in that time. But we should also expect 25% of our carbon emissions to be permanent, at least on the timescale of humans. That remaining fraction will need to be removed by the "weathering thermostat", which operates on timescales of millions of years.

Deglaciation

As a synthesis of some of what we've discussed, and a testament to the complexity of the Earth system, let's sketch part of the mechanism that led to the end of the last glacial maximum. This is a scenario where the ocean storage of carbon reversed, leading to drastic warming. We don't expect this to happen this century, but it's a cool story and shows the power of the ocean in controlling climate. Here are the steps:

1. An increase in summer solar radiation began to melt glaciers around 65°N
2. The influx of fresh water slowed the MOC, so that heat transport from southern high latitudes to northern high latitudes was slowed
3. Northern cooling, Southern warming
4. Melting of southern sea ice, southward shift of westerlies

5. Strong upwelling in southern ocean
6. Degassing of CO2

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