Chapter 10: Oceanic Circulation

Objectives:

- Ocean Structure
- Surface circulation --- Wind-driven Ocean Circulation
- Deep circulation --- Salt-driven circulation: Thermohaline Circulation
10.1 Ocean structure

Outline

- Size & shape of the ocean
- Continental drift
- Temperature structure of the ocean
- Surface currents
- Ocean’s role in global heat transport
Size & shape of the ocean

- Ave. depth \( \sim 4 \text{ km} \)
- Continental shelf: ave.width \( \sim 70 \text{ km} \), ave. depth at shelf break \( \sim 130\text{m} \)
- Further offshore => continental slope => abyssal plain (broad plain of deep ocean)
- Trenches (Submarine valley) : deepest 11 km.
- **Mid-ocean ridges:** new sea floor emerging
  => sea floor spreading & continental drift
- **Continental drift theory** (Wegener in 1912)
  - ~200 M yr. ago, Earth had only 1 super-continent **Pangaea**.
  - Pangaea splitted up => continents drifted apart.
  - Atlantic Oc. growing, Pacific shrinking.
Continental drift

a. About 225 million years ago

b. 135 million years ago

c. 65 million years ago

d. Present time
Temperature $T$

- **Vertical profile:**
  - Solar radiation absorbed within 100m of sea surface.
  - Wind $\Rightarrow$ surface mixed layer of 50-200m, ($T$ is nearly uniform).
  - **Thermocline** occurs between 200-1000m depth: $T$ decr. rapidly with depth.
  - Below thermocline, $T$ decr. very slowly to 0-3°C at oc. bottom.
Vertical temperature section in Atlantic

North

South
February sea surface temperature (SST)
August sea surface temperature (SST)
Surface currents

- **Gyres**: Large horizontal circulation cells.
Ocean’s role in global heat transport

- Oc. transports almost as much heat poleward as atm.:

Oc. dominates at low lat., atm. dominates at mid-high lat.
- **Heat capacity**: amount of energy needed to raise temp. of a unit mass by 1 °C.

- **Water has a high heat capacity:**
  - Temp. range over land many times that over oc., as heat cap. of water much larger than that of soils/rocks.
  - Oc. heat capacity \(~1600\) times of atm.
- Oc. has strong moderating effect on climate, e.g. coastal regions milder than inland.
- Large heat capacity => difficult to change oc. => oc. has long "memory" & major role in climate time scale, where atm. becomes "slave" to oc.
10.2 Wind-driven Ocean Circulation

-- Ekman motion and Ekman Spiral.
-- Upwelling & downwelling.
-- Geostrophic currents.
Inertial Motion

\[
\frac{du}{dt} = -\frac{1}{\rho} \frac{\partial p}{\partial x} + 2\Omega v \sin \varphi
\]

\[
\frac{dv}{dt} = -\frac{1}{\rho} \frac{\partial p}{\partial y} - 2\Omega u \sin \varphi
\]

\[
\frac{dw}{dt} = -\frac{1}{\rho} \frac{\partial p}{\partial z} + 2\Omega u \cos \varphi - g
\]
\[ \frac{\partial p}{\partial x} = \frac{\partial p}{\partial y} = 0 \]

\[ \frac{du}{dt} = 2\Omega v \sin \varphi = fu \]

\[ \frac{dv}{dt} = -2\Omega u \sin \varphi = -fu \]
Figure 9.1 Inertial currents in the North Pacific in October 1987 (days 275-300) measured by holey-sock drifting buoys drogued at a depth of 15 meters. Positions were observed 10-12 times per day by the Argos system on NOAA polar-orbiting weather satellites and interpolated to positions every three hours. The largest currents were generated by a storm on day 277. Note: these are not individual eddies. The entire surface is rotating. A drogue placed anywhere in the region would have the same circular motion. From van Meurs (1998).
Ekman Motion

- Ekman assumed a steady, homogeneous, horizontal flow with friction on a rotating Earth. Thus horizontal and temporal derivatives are zero:

\[
\frac{\partial}{\partial t} = \frac{\partial}{\partial x} = \frac{\partial}{\partial y} = 0
\]
- Coriolis Force = Wind stress
- Wind stress = tangential force on a unit area of ocean surface

\[ T_{xz} = \rho_w A_z \frac{\partial u}{\partial z}, \quad T_{yz} = \rho_w A_z \frac{\partial v}{\partial z} \]

\[ f v + A_z \frac{\partial^2 u}{\partial z^2} = 0 \]

\[ -f u + A_z \frac{\partial^2 v}{\partial z^2} = 0 \]
\( V_0 \) is the velocity (wind is blowing to the north) of the current at the sea surface.

Now let's look at the form of the solutions. At the sea surface \( z = 0 \), \( \exp(z = 0) = 1 \), and

\[
\begin{align*}
u(0) &= V_0 \cos(\pi/4) \\
v(0) &= V_0 \sin(\pi/4)
\end{align*}
\]

\[
a = \sqrt{\frac{f}{2A_z}}
\]
The current has a speed of $V_0$ to the northeast. In general, the surface current is 45° to the right of the wind when looking downwind in the northern hemisphere. The current is 45° to the left of the wind in the southern hemisphere. Below the surface, the velocity decays exponentially with depth:

$$[u^2(z) + v^2(z)]^{1/2} = V_0 \exp(az)$$
\[ u = V_0 \exp(az) \sin(\pi/4 - az) \]
\[ v = V_0 \exp(az) \cos(\pi/4 - az) \]
- **Nansen** (1890s) observed iceberg moving 20-40° to right of wind.

- **Ekman** (1905) solution has surface current at 45° to right of wind in N.Hem. (to the left in S.Hem.) (Coriolis effect).
**Ekman Mass Transports**

Flow in the Ekman layer at the sea surface carries mass. For many reasons we may want to know the total mass transported in the layer. The *Ekman mass transport* $M_E$ is defined as the integral of the Ekman velocity $U_E, V_E$ from the surface to a depth $d$ below the Ekman layer. The two components of the transport are $M_{Ex}$, $M_{Ey}$:

$$M_{Ex} = \int_{-d}^{0} \rho U_E \, dz, \quad M_{Ey} = \int_{-d}^{0} \rho V_E \, dz$$
The transport is perpendicular to the wind stress, and to the right of the wind in the northern hemisphere.
Application of Ekman Theory

Figure 9.7 Sketch of Ekman transport along a coast leading to upwelling of cold water along the coast. **Left:** Cross section. The water transported offshore must be replaced by water upwelling from below the mixed layer. **Right:** Plan view. North winds along a west coast in the northern hemisphere cause Ekman transports away from the shore.
Upwelling & downwelling

- Wind blowing alongshore can generate offshore Ekman transp. => upwelling of lower, cool nutrient-rich water => enhanced biol. productivity
- Onshore Ekman transp. => downwelling => poor biol. prod.
Along Equator, Easterlies => Ekman transport away from Eq. => strong upwelling along Eq.
- Upwelling under cyclones
- Downwelling under anti-cyclones.
- In N.Hem., surface current spirals to the right with incr. depth. Observ. wind driven layer (Ekman layer) is ~10-100m

- The depth-integrated mass tranport (Ekman transport) is at 90° to right of wind in N.Hem. i.e. wind balances Coriolis.
Geostrophic currents

- Tilt in sea level (SL) => pressure gradient => pressure (p) force. When p force is balanced by the Coriolis force => geostrophic current.
Gradual buildup of a geostrophic current:

- Low $p$ to High $p$
- $p$ force
- Coriolis force

Diagram showing the gradual buildup of a geostrophic current with force vectors from low pressure to high pressure.
- N.Hem.: low lat. easterlies, mid lat. westerlies
  => converging Ekman transport & high sea level (SL) at ~30°N

=> geostrophic currents.
Pressure gradient from SL tilt disappears by \( \sim 1000 \text{m} \) depth \( \Rightarrow \) geostrophic current only in top 1000m.

\[ p_1 \cdot \text{no } p \text{ force} \quad \cdot p_2 = p_1 \]
3 forces in upper ocean:

- wind stress, pressure gradient, Coriolis
- In Ekman layer (top 100m) mainly Coriolis balancing wind stress.
- 100-1000m: mainly Coriolis balancing pressure gradient => geostrophic current.
SL measurements from satellite

- Altimeter: measures return time of radar signal
  => distance to sea level
  => hills and valleys in the SL
  => geostrophic currents.
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10.3 Western Boundary Current

What causes intense western boundary currents?
**Vorticity:** describes the tendency of a fluid to rotate.

- Clockwise rotation => negative vorticity
- Counterclockwise rotation => positive vorticity

Vorticity is an attribute of rotation. Any rotation generates vorticity.
The vorticity generated by the earth rotation is called **planetary vorticity**. Any object in a place between the equator and poles has vorticity.

Planetary vorticity = \( f \) (Coriolis force).

The other rotations rather than the earth rotation also generate vorticity, called **relative vorticity**.
- Vorticity measures the intensity of rotation. 
  more intense rotation $\iff$ larger
positive vorticity

negative vorticity

Clockwise vorticity

Counterclockwise vorticity

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- Coriolis effect max. at poles, 0 at equator.
- If Coriolis constant with latitude $\Rightarrow$ no western intensification.
Stommel: \textbf{Coriolis force constant}

\textbf{streamlines} \hspace{1cm} \textbf{Coriolis force varies linearly with latitude} \hspace{1cm} \textbf{sea level height}

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10.4 Deep circulation --- Thermohaline

- Composition of “salt” in oc.
- Distr. of salinity in oc.
- What affects density of sea water
Salinity:

measured in terms of the proportion of dissolved salt to pure water.

(unit: g salt /kg seawater)
Salinity S

- Ave. concentration of salt in sea water (i.e. S) is ~3.5%.
- Until early 1980s, S expressed in parts per thousand, 3.5% written as 35 o/oo. The o/oo symbol now discarded.
- Major constituents of S:

<table>
<thead>
<tr>
<th>Component</th>
<th>%</th>
</tr>
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<tbody>
<tr>
<td>Chloride, Cl⁻</td>
<td>55.0</td>
</tr>
<tr>
<td>Sodium, Na⁺</td>
<td>30.6</td>
</tr>
<tr>
<td>Sulphate, SO₄⁻²</td>
<td>7.7</td>
</tr>
<tr>
<td>Magnesium, Mg⁺²</td>
<td>3.7</td>
</tr>
<tr>
<td>Calcium, Ca⁺²</td>
<td>1.2</td>
</tr>
<tr>
<td>Potassium, K⁺</td>
<td>1.1</td>
</tr>
<tr>
<td></td>
<td>99.3</td>
</tr>
</tbody>
</table>
How to measure salinity?

• Get rid of the water and weigh the salts left behind. Unreliable.
• Higher $S \Rightarrow$ more ions $\Rightarrow$ incr. elec. conductivity.
  Since mid-1960s, measure conduc. to get $S$.
• $S$ measured by a CTD (Conductivity-Temperature-Depth) probe.
As the CTD instrument is lowered through the water (or as it sits still at a given point), measurements of conductivity, temperature and depth are recorded continuously.

CTD instruments measure three important quantities directly - conductivity, temperature and pressure. By measuring conductivity $\Rightarrow$ salinity.
Distr. of sea surface salinity (SSS)

- River runoff => low SSS near coast.
- Melting snow & ice => lower SSS at high lat.
- Pac. Oc. much less saline than Atl. Oc.
Lat.distr. of SSS follows Evap.-Precip. (E-P).
- **Halocline** = region of strong change of S with depth, at ~ 200m-1000m.
- **Typical halocline**: Low lat., S decr. with depth
  High lat., S incr. with depth
- In open oc., density \( \sim 1.022-1.030 \text{ g/cm}^3 \).
- Density determined by T, S & pressure p.
- At mid & low lat., density profile approx. T profile.
Thermocline => **pycnocline** (where density changes rapidly with depth).
- **TS diagram:** T-S-Density relation.
- **At low T,** changing T has little effect on density.
Bottom water formation

- surface water: 0-500 m deep,
- intermediate water: 500-1500 m,
- deep water: 1500-3500 m,
- bottom water: > 3500 m

Q: What conditions needed to form bottom water?
   (a) intense cooling, or
   (b) incr. in S. Usu. both (a) & (b) required.

- Polar regions during winter: cooling and windy cause dense water (strong wind will evaporate water and leave salt behind) (cool and salty water);
- When sea water is frozen into ice, much of salt expelled into surrounding water, since ice can’t contain the salt. So the water underlying the sea ice is very relatively salt.
There are two important regions to form deep water. North Atlantic and Antarctic because they are very cold, and windy. The dense water masses that sink into the deep basins are formed in quite specific areas of the **North Atlantic** and in the **Southern Ocean**.
By contrast in the **Weddell Sea** off the coast of **Antarctica** near the edge of the ice pack, the effect of wind cooling is very intense. The resulting **Antarctic Bottom Water** (ABW) sinks and flows north into the Atlantic Basin. The flow into the Pacific is blocked.

Bottom water formed off Antarctica, mainly in Weddell Sea.

**Antarctic Bottom Water (ABW)** is densest water in open oc.

In the Norwegian Sea evaporative cooling (large wind leading to strong evaporation and in turn leading to large salinity), is predominant, and the sinking water mass, the North Atlantic Deep Water (NADW), fills the basin and moves southwards. It then flows very slowly into the deep abyssal plains of the Atlantic, always in a southerly direction.
- N. Pacific is too low in S to form bottom water. Cooling in high lat. => intermediate water.
- **Q:** Where does the Pac. gets its bottom water?
  
  Its bottom water (the *Common Water*) is a mixture of NADW & ABW, introduced into the Pac. by the Antarctic Circumpolar Current.
NADW flows southward through the Atlantic Oc. And joins with Antarctic Circumpolar Current, which flows around Antarctica. There the NADW and ABW combine and circle the continent. They then proceed to branch off into the Indian and Pacific Oceans.
Thermohaline circulation

Thermohaline circulation: The density of seawater is controlled by its temperature \((\text{thermo})\) and its salinity \((\text{haline})\), and the circulation driven by density differences is thus called the thermohaline circulation. The thermohaline circulation is sometimes called the ocean conveyor belt, the global conveyor belt, or, most commonly nowadays, the meridional overturning circulation.

Top 1 km dominated by wind-driven oc. circ., below 1 km, thermohaline circ. dominates.
Originally the deep water is formed in North Atlantic, near Greenland, Iceland and Norwegian sea (NADW). The NADW sinks into bottom and then further moves southward. The NAWD will move to Antarctic region and merge with ABW (Antarctic bottom water), and move northward to arrive at the North Pacific. Meanwhile, the surface current near the western Pacific ocean moves southward in the form of gyre, and further cross Indian ocean and back to Atlantic ocean to replace water there sinking into bottom.

So, the thermohaline circulation includes a deep ocean circulation from the North Atlantic Ocean to the North Pacific to bring deep water (salty and cold) into Pacific Ocean; and a surface current from the North Pacific to North Atlantic ocean. Both circulations act to make the water mass conservation.
The effect of Thermohaline circulation on climate

(1) THC transports heat from the south to North to warm the North Atlantic and Europe.

(2) adjust the low latitude climate too by transporting surplus heat
Change in annual temperature 30 years after a collapse of the thermohaline circulation
Tracers

- Tracers (e.g. T, S, oxy., other chemicals) used to infer movements of water.

- T & S changed only at sea sfc.
  => "conservative tracers" after leaving sfc.
  (S changed by precip.-evap., river runoff, ice formation).

- Oxygen saturation when in contact with atm.
  Just below sfc., oxy. incr. from biol. productivity.
Atlantic section

Temperature

Salinity

Oxygen (ml/l)