Chapter 10. Thermohaline Circulation

Main References:


Other references:


10.1 The Thermohaline Circulation (THC): Concept, Structure and Climatic Effect

10.1.1 Concept and structure
The Thermohaline Circulation (THC) is a global-scale ocean circulation driven by the equator-to-pole surface density differences of seawater. The equator-to-pole density contrast, in turn, is controlled by temperature (thermal) and salinity (haline) variations. In the Atlantic Ocean where North Atlantic Deep Water (NADW) forms, the THC is often referred to as the Atlantic Meridional Overturning Circulation (AMOC). Changes in surface heat and salinity fluxes can affect the pole-to-equator density differences, and thus may affect the AMOC. Different from the wind-driven ocean circulation, which is generally in the upper 1–2km of the ocean, the AMOC extends to the very deep ocean.

The general idea is that near the equator, the ocean receives excess radiative heating and thus heats up the ocean; near the pole, the ocean has heating deficit because the outgoing longwave radiation is larger than the downward shortwave radiation and thus cools the ocean (Figure 1, left). As a result, the equator-to-pole density gradients are set up. The density gradients are particularly strong in the North Atlantic (Figure 1, right), because the Atlantic basin extends farther north than the Pacific and Indian Oceans, where
wintertime storms can increase the surface cooling by draining the latent and sensible heat fluxes. In addition, sea ice formation can eject salt, and thus increase sea surface density.

Figure 1. Left: Schematic diagram showing the shortwave and longwave radiative fluxes as a function of latitude. (right) Mean Atlantic Sea Surface Temperature (SST). In addition to radiation deficit, winter storms bring cold and dry air to the North Atlantic, which can increase the cooling by draining latent and sensible heat from the ocean. Sea ice formation ejects salt, which will increase the sea surface density. Both processes act to enhance the pole-to-equator surface density contrast.

An over-simplified, zonally mean picture of the THC is shown in Figure 2 (top-left). With cooling near the pole and heating near the equator, a meridional overturning circulation forms, with cold water sinks at high latitude, flows equatorward in the deep ocean, upwells to the surface in the tropics-to-midlatitudes, and the warm surface water flows to the pole to close the MOC. In fact, the observed temperature and salinity along 30°W longitude of the Atlantic indeed show strong SST gradients from the equator to pole, and the gradients are especially strong from 30°N-50°N. The cold surface water extends all the way down to the deep ocean at high latitude, and this downward extension can also be seen in salinity field (Figure 2, top-right and bottom-right panels). Note that the situation is not as simple as the idealized picture shown in the top-left panel of Figure 2: there are fresher
Antarctic Intermediate Water (AAIW) introducing into the the North Atlantic, and the NADW flowing out to the southern ocean (bottom-right of Figure 2). A zonal-mean, more realistic view of the AMOC is shown in bottom-left panel of Figure 2, with warm surface water flowing northward, and cold water sinks in the north and forms the NADW, which then flows southward into the Southern Ocean. Meanwhile, the Antarctic Bottom Water (AABW) flows into the Atlantic at the bottom.

**MOC: complex structure**

**Schematic**

**Observations**

Figure 2. Left: schematic diagram showing the zonal mean structure of AMOC; Right: observed latitude-depth section along the 30W of the Atlantic for temperature (top-right) and salinity (bottom-right).
9.1.2 Climatic effect
Given that the warm, upper-ocean water flows northward into the North Atlantic, and the cold NADW flows out (Figure 3), the AMOC transports net heat into the North Atlantic. This northward net heat transport is shown in Figure 4, which is calculated from the NCEP reanalysis data (Trenberth and Caron 2001). Different from the Pacific where the excessive heat gained in the tropics is transported poleward into both the North and South Pacific, and the Indian Ocean where the excessive heat in the tropics is transported southward (since the Indian Ocean is bounded by lands in the North), the Atlantic transports heat “northward” in both the southern and northern hemispheres. This marked contrast is due to the strong AMOC in the Atlantic, which plays a crucial role in transporting heat toward the north pole. In the Pacific and Indian Oceans, however, the heat is transported out of the tropics primarily by the wind-driven, shallow meridional overturning circulations, which are referred to as the “subtropical cells (STCs)”.

Figure 4. Zonal-mean meridional heat transport for the global ocean (black), Pacific (green), Indian (red) and Atlantic (purple). Adapted from Trenberth and Caron (2001).
An important question is: What will happen if the AMOC slows down or collapses? Evidently, a slowdown AMOC will reduce northward heat transport, and thus cool the North Atlantic. In fact, a collapsed AMOC was suggested to be the major cause of the Younger Dryas event, which is a global-scale cooling event that occurred during ~12,800 ~ 11,500 yrs before present.

Global coupled atmosphere-ocean model experiments show that a slowdown AMOC generates large cold SST anomalies in the north and warming SST anomalies in the south Atlantic (Figure 5). These SST anomalies induce atmospheric circulation changes in the coupled climate system, thereby affecting global climate (such as the ITCZ, sea level change along the US coast, etc.). The slowdown AMOC will also affect CO$_2$ transport into the deep ocean.

![Figure 5. SST anomalies in responding to a 1sv-freshwater forcing in the North Atlantic, which generates a weakened AMOC. Adapted from Stouffer et al. (2006).](image_url)

### 10.2 Mechanisms

Since the AMOC is a global-scale ocean circulation and its variability has large impacts on the ocean and climate, extensive observational, theoretical and modeling studies exist to understand its variability and mechanisms. The mechanism understanding, however, is far from complete, due to the complex nature of the AMOC dynamics. The AMOC is forced by surface buoyancy flux, and is associated with deepwater formation and wintertime convection. Meanwhile, it is affected by internal diffusion, northward transport of upper-ocean water from the Southern Ocean. It is also influenced by winds, basin geometry, bottom topography, advection of heat and salinity, and small-scale processes. Given this shear of complexity, a hierarchy of models approach is useful,
which can extract the major physical processes that generate the AMOC, and then adding other processes in an orderly manner.

Here, for this class, we introduce the fundamental processes that are essential for producing the AMOC, using a simple 2-layer ocean model. We will also show some corresponding solutions from an Ocean General Circulation Model (OGCM), to verify the 2-layer model performance, and confirm that the major physical processes identified in the 2-layer model have realistic meanings. The model basin will be a rectangular box with 0-40° in longitude and 0-60°N in latitude and flat ocean bottom, which represents the North Atlantic basin but ignores the slanted side boundaries and bottom topography.

The following hierarchy of experiments using the 2-layer model will be discussed:

1. Heat flux $Q(y)$ forcing, which is a function of only latitude ($y$) and independent of longitude ($x$); the ocean is inviscid;
2. Add diffusion to (1);
3. Allow $Q$ to vary with longitude, so $Q$ varies with both $x$ and $y$ - $Q(x,y)$;
4. Add an inflow at the southern boundary to represent the inflow from the Southern Ocean.

Note that this set of experiments is at the base of the hierarchy toward a more comprehensive understanding of the AMOC dynamics. Our goal is to introduce the major physical processes that are essential for producing the AMOC. The key questions are:
Can the “localized wintertime deep convection plumes” drive the AMOC? Is the sinking branch of the AMOC associated with large-scale oceanic dynamical adjustment processes? Since the pole-to-equator density contrast generates zonal geostrophic flow, how can the system develop meridional circulation (AMOC)?

(1) **Heat flux $Q(y)$ forcing, with no diffusion**
Here, we include $Q(y)$ effect by relaxing the model SST to a prescribed temperature profile $T^*$ (red line in Figure 6), which equals $T_s=23^\circ C$ from 0-30N ($y_1$), linearly decreases to 3$^\circ C$ at 50N ($y_2$), and then keeps 3$^\circ C$ north of 50N. Initially, in responding to the $T^*(y)$, there is a meridional pressure gradient in layer 1, and the response across the interior ocean is zonal current $U_1$ ($V_1=0$) that flows eastward within the “$T^*(y)$ gradient” region. Note that the large $T^*$ gradient region of 30N-50N is consistent with the observations (Figure 2, top-right). However, $T$ gradients exist south of 30N in the observations, even though the gradients are weaker. Since there is no wind forcing, the ocean response is baroclinic, and thus layer 2 mirrors layer-1’s currents. Along the eastern boundary there is a convergence of layer-1 water due to $U_1$ that depends $h_1$. [Conversely, $U_1$ drains layer-1 water from the coast along the western boundary, in order to keep $h_1 \geq h_{\text{min}}$.] Along the eastern boundary, coastal Kelvin waves radiate northward, and after their passage the coastal layer thickness adjusts to ensure that there is no flow into the coast. So $h_1$ deepens near the eastern boundary within the Rossby radius and intersects the bottom at latitude $y'$. In responding to the $h$ changes, there is northward
flow in layer 1 and southward flow in layer 2 (Figure 6, top-right). The deepened h1 signal then propagates westward via baroclinic Rossby waves. Since the ocean is inviscid, he propagates unchanged across the basin as an interfacial front (bottom left). Everywhere east of it, h1 is adjusted to the eastern boundary value, he (which is h1 along the eastern boundary), and the ocean is in a state of no motion there. After Rossby waves propagate to the western boundary, the ocean reaches a state of balance with no flow everywhere in the basin (Figure 6, bottom right).

This oceanic adjustment process with deepened h1 by radiating coastal Kelvin and Rossby waves essentially describes the “deep water formation” or “descending branch” of the AMOC. Without diffusion, however, there is “no AMOC”! This means that even if we have “deepwater formation”, or wintertime convection, water will comes up somewhere when cold water falls down. So, only convection itself cannot form AMOC.

(2) Heat flux Q(y) forcing + diffusion (y<2)

By adding diffusion in the T* gradient region, a model AMOC is generated, with upwelling occurring all in the western boundary region (Figure 7). In this case, the first two stages of oceanic adjustments are the same as we discussed above. Therefore, we only discuss stage 3 and stage 4. While Rossby waves propagate westward, mixing damps them, and meanwhile mixing also thickens h1 (Figure 7, right panel). Due to the Rossby waves' damping, zonal pressure gradients are set up, which sustain northward geostrophic V1. Steady state solution is an AMOC that is closed within the T* gradient region (right panel of Figure 7). A final state 5 is also needed for the ocean to reach equilibrium state, which involves multiple reflections of Rossby waves from the eastern boundary (arrive at the western boundary - coastal Kelvin waves to the EQ - EQ Kelvin waves to the eastern boundary – coastal Kelvin waves northward – then Rossby waves westward). These processes occur until upwelling balances downwelling. The equilibrium state involves a western boundary current that extends all the way to the equator (not shown).

Figure 7. (left) Stage 3 and stage 4 of oceanic adjustment for the viscid case but with diffusion only applied within the T* gradient region.
With basin-wide diffusion, an AMOC is generated with upwelling occurring throughout the interior basin (Figure 8). The western boundary layer V1 transport, interior and total meridional transports are shown in Figure 9, which are obtained by finding analytical solutions. A closed AMOC is formed, which resembles the Stommel and Arons (1960) solution and Kawase (1987) solution.
Figure 9. Layer 1 thickness (upper panel) and meridional transports (bottom) for the 2-layer model with diffusion everywhere in the basin.

(4) Heat flux $Q(x,y)$: strongest cooling in the northwest basin

Note that one unrealistic aspect of our above solution is that $Q$ is independent of $x$, whereas in observations cooling occurs in the northwest basin, such as the Labrador Sea, Greenland, Iceland, and Norwegian seas. To simulate the more realistic cooling, we relax the SST to $T^*(x,y)$ which obtains the coldest value in the northwest. Isotherms in the cooling band tilt poleward toward the east. The dynamics are essentially unchanged, although the solutions are. During the spin-up, Kelvin waves till adjust the eastern-boundary layer thickness and Rossby waves still carry the eastern-boundary stratification into the interior ocean. In this case, however, he also extends along the northern boundary, and the Rossby waves propagate westward and southward within the cooling band, allowing the northern-boundary coastal structure to propagate into the interior ocean as well. The steady, interior solution is shown in Figure 10, in which downwelling (or descending branch of the AMOC) occurs throughout the northern boundary.
Figure 10. (Top) Horizontal map of layer-1 thickness h1 (white contours), transport (arrows) and w1 (upwelling/downwelling, color) for the ocean interior from the 2-layer model. Ignore the detailed marks for A, B, C D, etc. (Bottom) similar to the top but from the OGCM numerical solution with a western boundary.

(5) With inflow from the Southern ocean
By considering the inflow from the southern ocean, the oceanic thermocline depth adjusts, and for inviscid case, all deepwater is supplied by the southern ocean. In reality, however, the ocean is viscid, and part of the deepwater is compensated by interior upwelling (Figure not shown).

In summary, the large-scale oceanic dynamic adjustments are important for the descending branch of the AMOC. The adjustment process, including Rossby wave damping, is important for setting up zonal pressure gradients and form a northward flow in the upper ocean (the northward branch). The upwelling in the ocean interior (upwelling) and inflow from the Southern ocean is important for closing the AMOC. The
deep southward flow is important for the NADW to flow out to the Southern Ocean. In reality, the AMOC is closed by the upwelling in the world’s oceans.