Lecture 8: Thermohaline Circulation and Salinity Effects

8.1 Preliminaries

The thermohaline circulation (THC) is driven by surface fluxes of heat (changing the temperature – “thermo”) and freshwater (changing the salinity – “haline”), which combined change water density and hence pressure. The terms “thermohaline circulation” and “buoyancy-driven flow” (buoyancy = gravitational acceleration times density anomaly) are, strictly speaking, synonymous. But there has been a tendency to use the term THC in a more restricted sense, for that part of the ocean circulation associated with convection and sinking, upwelling from depth, and the horizontal flows feeding these vertical motions. There is no reason to prefer one expression to the other, except perhaps when one wishes to emphasise that temperature and salinity interact very differently with the atmosphere (see Lecture 9). In that case, referring explicitly to temperature and salinity through the term “THC” is in part a reminder of these different interactions. A more important issue of notation arises because ocean dynamics are nonlinear, making it impossible rigorously to separate wind-driven from buoyancy-driven circulations. For simplicity, we will nevertheless often pretend that they are separable.

8.2 Deep western boundary currents

What are the dynamics controlling the global THC, as depicted, for example, in Gordon’s (1986) **cartoon**? Even non-oceanographers know that the surface flows are concentrated in western boundary currents, which are poleward in the subtropics and middle latitudes. In the North Atlantic, much of the net northward near-surface mass transport of the THC (e.g., **Fig.** 11, Macdonald, *Prog. Oceanogr.*, **41,** 281-382, 1998) occurs in the Gulf Stream, which is thus a prime example of a current that is both wind- and buoyancy-driven.

It is less well known that the deep flow is likewise concentrated near the western boundary. After it was discovered in the early 19^{th} century that deep water was cold even at low latitudes and hence had to originate from high latitudes, the deep equatorward flow was thought to spread over the entire basin. But Stommel and Arons (1960; see Warren, 1981, for a beautiful discussion of the history) predicted that strong boundary currents had to exist in the deep ocean as well. Hence, the flow of North Atlantic Deep Water occurs mainly in such a deep western boundary current (DWBC).

We now go through a very simple rendition of the Stommel-Arons theory (**Sketch**). Assume that the ocean is flat-bottomed and consider the vertically integrated flow in a bottom layer. Assume horizontally uniform upwelling from this layer, compensating strong, localised downwelling at the convection sites. (N.B.: All these assumptions have since been severely discredited. Ocean bottom topography is important, upwelling – probably associated with regions of vigorous mixing – might be highly localised rather than widespread, and downwelling is unlikely to occur at the convection sites. Still, the prediction of DWBC's by the Stommel-Arons theory is striking, and the vast majority of oceanographers consider it the standard theory of deep circulation).

Starting from geostrophy (we use Cartesian coordinates for simplicity),

$$-fv = -\frac{1}{\rho_0}\partial_x p\,, \tag{8.1}$$

$$fu = -\frac{1}{\rho_0}\partial_y p\,, \tag{8.2}$$

cross differentiation and subtraction of the equations yields

$$-f\left(\partial_x u + \partial_y v\right) - v\frac{df}{dy} = 0\,. \tag{8.3}$$

With the continuity equation

$$\partial_x u + \partial_y v + \partial_z w = 0 \tag{8.4}$$

and the definition $\beta \equiv df/dy$, this gives

$$\beta v = f\partial_z w\,, \tag{8.5}$$

which is called the planetary (or linear) vorticity equation. Integrating (8.5) over the bottom layer gives (the flat bottom implies that $w = 0$ there)

$$\beta V = fw, \tag{8.6}$$

which says that upwelling out of the deep layer implies *poleward* horizontal flow (toward the source!), in contrast to the 19th-century expectation. Making the standard "beta-plane" approximation, $f = \beta y$, leads to

$$\partial_y V = w \tag{8.7}$$

from (8.6), so the poleward flow even increases with latitude. If planetary vorticity conservation holds at all longitudes, (8.7) implies for the zonal integral across the entire basin that

$$\int_0^L \partial_y V dx = \int_0^L w dx\,. \tag{8.8}$$

But the vertically integrated continuity equation leads to a different conclusion. From (8.4),

$$\partial_x U + \partial_y V + w = 0\,. \tag{8.9}$$

Integration across the basin yields, noting that the zonal flow vanishes at the eastern and western boundaries if these follow longitude lines,

$$\int_0^L \partial_y V dx = -\int_0^L w dx\,, \tag{8.10}$$

the exact opposite of (8.8)! Both equations cannot be right, and since mass conservation is correct to high accuracy (here with the very sensible interpretation that the upwelling is fed from the decreasing northward transport), we conclude that (8.8) cannot be true everywhere. A boundary current must exist with a dynamical balance different from the linear vorticity relation. This boundary current must supply both the interior increase in V with latitude and the upwelling out of the deep, which are equal if the width of the boundary current is much less than L. That the boundary current must be on the western side follows from arguments similar to those used for the westward intensification of surface currents, and we will not go into any detail here. For example, considering the vorticity balance suggests a western boundary current to balance the vortex stretching by upwelling out of the bottom layer.

If one assumes a mass source in the northwestern corner of a basin, a southward boundary current emanates, which weakens as it progresses southward, due to “leakage” to the east (**Sketch**). Depending on the strength of the source, the DWBC loses all its mass at some latitude or reaches the Antarctic Circumpolar Current (ACC). Half of this mass lost upwells over the basin, while the other half reaches the northern boundary, from where it recirculates (according to unspecified dynamics) to the northwestern corner.

The details of the ensuing picture (**Fig.**, Kuo & Veronis, after Warren 1981) depend on the assumed distribution of mass sources, but two conclusions are robust:

i) Flow away from the source occurs in the DWBC only; all interior flow is poleward.

ii) Cross-equatorial flow occurs only in the DWBC.

How well does the Stommel-Arons theory describe the real ocean? Generally, one should be very critical, but the existence of DWBCs is a robust phenomenon: We see them in the real ocean. **Fig.:** Dickson et al. (*Nature*, **344**, 848-850, 1990) made direct current measurements off the East Greenland coast. The DWBC shows up as a very fast (<30 cm/s) and narrow (ca. 40 km) flow of very high-density water “hugging” the western topography. This water is also very cold (**Fig.:** Smethie and Swift, 1989, *J. Geophys. Res.,* **94**, 8265-8275). Equatorward in the Atlantic, the DWBC is most easily identified through the Chlorofluorocarbons (CFCs), which are purely anthropogenic and show waters that recently were in contact with the atmosphere (**Fig.** 8 from Smethie, 1993, *Prog. Oceanogr*., **31**, 51-99). In the South Atlantic, there is a very clear signature of the DWBC in all properties: High temperature, high salinity, high oxygen, low silica (**Fig.** 1.10 from Warren, 1981).

Curiously at first, the DWBC now is a buoyant, rather than dense, anomaly when compared with the ambient water at the same latitude. This has to be so from thermal-wind considerations, meaning that the deep zonal pressure and hence density gradients must change sign at the equator as the DWBC crosses the equator. But even a rudimentary dynamical explanation is very complicated (Marotzke and Klinger, *J. Phys. Oceanogr.*, **30**, 2000, in press).

The high salinity and low silica characteristic of North Atlantic Deep water can be found in the South Indian and South Pacific Oceans to about the equator; then the trace is lost.

8.3 Salinity effects

Seawater consists of freshwater and salt. (The dissolved salts have a nearly universal mixing ratio; for physical purposes, it is sufficient to know the total salt mass). Temperature and salinity have influences on density that are of the same order of magnitude, in contrast to the atmosphere where water vapour has a noticeable influence on density only in the tropics.

The thermal expansion coefficient, α, is defined as

$$\alpha \equiv -\frac{1}{\rho}\frac{\partial \rho}{\partial T}, \qquad (8.11)$$

and is a function of temperature, salinity, and pressure (see Gill, 1982, Table A3.1). Unlike freshwater, seawater has a density maximum below the freezing point (**Figs.** 2.07 and 2.08, Dietrich et al., *Allgemeine Meereskunde*, 3rd ed., Bornträger, Berlin), but this unattainable density maximum can be felt at low temperatures where α is very small. At the surface, α varies by one order of magnitude, from $0.25\times10^{-4}K^{-1}$ at -2°C to $3.4\times10^{-4}\,K^{-1}$ at 31°C. When we assume an intermediate value of $\alpha(13°C)=2\times10^{-4}\,K^{-1}$ as representative, the observed temperature range of 25°C gives a thermally-induced density range of $\Delta\rho_T$=5 kg m^{-3}.

The haline expansion coefficient, β, is defined as

$$\beta \equiv \frac{1}{\rho}\frac{\partial \rho}{\partial S} \qquad (8.12)$$

and varies by only about 10%, so a value of $\beta=8\times10^{-4}$ is appropriate for seawater. With a salinity range of perhaps 2.5, this leads to a salinity-induced density range of $\Delta\rho_S=2$ kg m^{-3} – less than, but of the same order of magnitude as, $\Delta\rho_T$.

At high latitudes, however, where α is very small, salinity fluctuations can dominate density fluctuations. In particular, surface salinity decides whether in a given winter deep convection occurs (see **Fig.** 6 from Dickson et al., *Prog. Oceanogr.*, **20**, 103-151, 1988): Typically, cool and fresh water overlays warm and saline water. The surface salinity then decides about whether winter cooling makes water dense enough to overturn convectively, or whether cooling even to the freezing point leaves surface water too buoyant. In the latter case, see ice forms, which insulated the ocean very effectively against further heat loss, and convection is suppressed. (N.B.: If ice gets exported, for example by wind drift, the continuously forming new sea ice might eventually leave enough salt behind in the surface layer that convection occurs later in the winter). As a rule, high surface salinity is needed for convection to occur regularly; the North Pacific is so low in salinity that is never convects to great depths (**Fig.,** SSS from Levitus, 1982, atlas). The contrast in sea surface salinity determines the different roles of Atlantic and Pacific in the global THC (Warren, 1983).

But why are the two oceans so different in surface salinity? To understand this, we must ask “what if” questions, for example, “What would it take to make the Atlantic look like the Pacific, and vice versa?” This requires the use of models, because “The exchange of letters in *Nature* … shows how futile simple verbal arguments can be in discussing such issues. The reader with a morbid interest in fallacious verbal theories may find it entertaining to look over the work of the English eccentric William Leighton Jordan…” (From footnote 3 in Robinson and Stommel, *Tellus*, **11**, 295-308, 1959). The next lecture deals with how to force models of the THC. After we understand that, we will be ready to simplify the models to the extent that a conceptual model can tell us why oceans can behave as differently as North Atlantic and North Pacific.