Chap 10 Thermohaline Effects

10.1 The deep circulation

The deep circulation is much less well known and less well described dynamically than the upper-layer circulation.

(1) Because of the difficulty of making direct measurements of velocity in the deep water.

(2) Most of the information on the deep circulation has been deduced from observations of distributions of water properties such as temperature, salinity, oxygen and trace elements.

(3) The property distributions suggest that the main sources of deep water are in the North and South Atlantic. For the North Atlantic, the major source of deep water is overflow from the Norwegian Sea over the sills between Greenland and Scotland.

(4) There is some evidence of sinking due to winter cooling in the Labrador Sea south of Greenland but it is probably very localized in both space and time. In the South Atlantic, the main source is probably the Weddell Sea where sinking results from density increase due to freezing out of ice. Other probable sources around Antarctica are the Ross Sea in the South Pacific and off the coast around 50°E and 140°E.

(5) The processes in both north and south are thermohaline and, of course, seasonal. In addition, there is evidence that formation of deep water is intermittent even during the cooling season.

(6) Contributions of thermohaline origin at mid-depth from the Mediterranean to the Atlantic and from the Red Sea and Persian Gulf to the Indian Ocean. These are waters rendered dense by evaporation at the surface, which then sink and flow out of the sea into the neighboring oceans.

Theory for thermohaline circulation was first proposed by Stommel.

(1) The depth of the thermocline at any locality remains substantially constant. Because in low latitudes there is a net annual inflow of heat through the surface into the water, the upper warm layer, and its boundary the thermocline, should deepen with time.

(2) As this deepening does not happen, some mechanism must be opposing the tendency, and Stommel suggests that this mechanism is slow upward flow of

cool deep water.

(3) Continuity requires that the sinking water in the North and South Atlantic must be balanced by rising, and Stommel suggests that while the sinking is very localized, the rising is spread over most of the low and middle latitude areas of the oceans shown in Fig. 10.1.

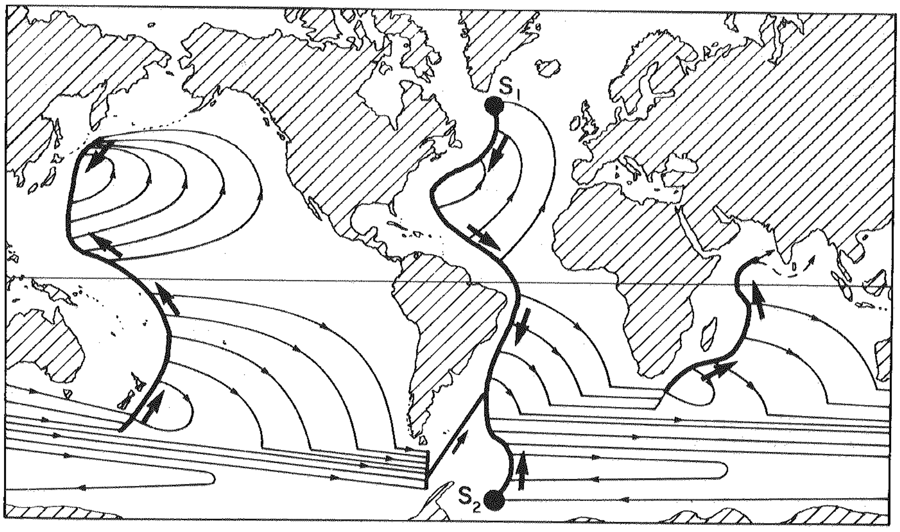

Fig. 10.1 Model for deep ocean circulation. (After Stommel, 1958)

(4) The sinking regions (S_1, S_2) are shown feeding relatively intense western boundary currents (required by conservation of vorticity in a situation in which the relative vorticity, ζ, is known to be small in the interior).

(5) Outward from these flow gentler geostrophic currents into the bodies of the oceans to supply the slow upward flow to maintain the thermocline depth constant.

(6) In the interior, upward motion causes D to increase; water moves poleward and the magnitude of f increases; ζ stays small. To get back south or north as necessary with ζ small requires input of vorticity of the appropriate sign. This input may be achieved with a strong flow and shear on the west again, just as for the upper layer circulation as discussed in Section 9.12.

(7) Fig. 10.2 shows that the strong flow and shear must be on the west rather than on the east when the return boundary flow is to the south, since southward flow requires input of negative vorticity to keep ζ small.

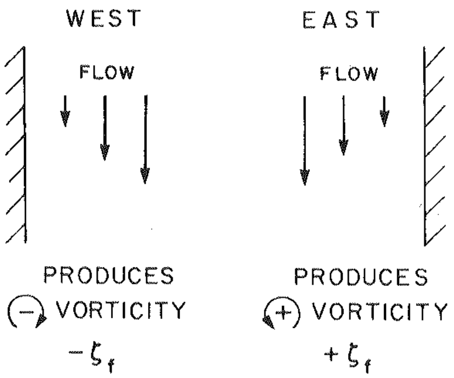

Fig. 10.2 Relative vorticity supplied by velocity shear at west and east sides of an ocean.

(8) Observations by Warren (1977) indicate an additional feature of the deep flow in the Indian Ocean. Earlier observations had shown the northward flow expected from Stommel's model along the west side of the ocean off Madagascar while the new observations showed a northward flow along the Ninetyeast Ridge which rises to about 4000 m depth along that meridian which is to the east of the center of the Indian Ocean. The inference is that a western boundary-type flow can be associated with the mid-ocean ridges, if high enough above the bottom, as well as with the western boundary itself.

(9) The deep currents are opposite in direction to the surface flows. The latter may be expected to be stronger to maintain continuity. The sinking water must be replaced by water which has come up through the thermocline and returns in the upper layer.

(10) Conservation of vorticity again requires western boundary currents in the return flow. Above the thermocline the upward flow causes D to decrease. To keep ζ small the flow is equatorward, i.e. opposite to the deep flow and likewise to conserve vorticity the western boundary flows in the upper layer will be opposite to those in the lower layer.

(11) The strong Gulf Stream to the northwest in the upper layer is consistent with the strong southwest flow in the deep water from S_1, while the less strong Kuroshio is associated with the weaker deep flow in the Pacific, although the thermohaline flow does still enhance the Kuroshio in the surface region compared with the purely wind-driven values. (Note that there is no large source of deep water in the Pacific. There is now believed to be some outflow from the Ross Sea in the Antarctic but the volume appears to be much smaller than that from the Weddell Sea.)

(12) The relatively weaker southward Brazil Current in the upper layer in the South Atlantic is consistent with the southward deep flow below it.

(13) Stommel suggests that the flow in the deep current under the Gulf Stream is about 30 Sv. The equal surface return flow would then almost double the Gulf Stream transport associated with wind driving.

(14) A limited number of deep current measurements with Swallow floats have in some cases supported the model and in some cases opposed it.

10.2 Equations for salt and temperature (heat) conservation

The differential equations for salinity and temperature are

$$\frac{dS}{dt} = \kappa_s \nabla^2 S \tag{10.1}$$

$$\frac{dT}{dt} = \kappa_T \nabla^2 T + Q_T \tag{10.2}$$

where κ_S and κ_T are molecular kinematic ***diffusivities*** ($m^2 s^{-1}$) for salt and for temperature, respectively. κ_T is about $\nu/10$ and κ_S is about $\nu/1000$. It has been assumed that κ_S and κ_T vary with position slowly enough that such variations can be ignored. A similar function to Q_T for the salinity equation, Q_S, is not required because processes affecting salinity occur only at boundaries, e.g. river inputs, effects of freezing or the difference between evaporation and precipitation.

A fluid element can change its salinity by molecular processes. When these elements are in turbulent flow, the turbulence, in "stirring" the fluid, makes the instantaneous property gradients very large and greatly increases the rate of change of properties of the elements compared with the rate in a non-turbulent fluid with a comparable mean gradient. In the averaged equations the mixing is described by the Reynolds stresses for momentum.

10.3 Equations for the average salinity and temperature

We adopt Reynolds' approach of splitting the total quantities into mean and fluctuating parts: $S = \overline{S} + S', u = \overline{u} + u'$, etc., and take the average of the equation.

$$\frac{\partial \overline{S}}{\partial t} + \overline{u}\frac{\partial \overline{S}}{\partial x} + \overline{v}\frac{\partial \overline{S}}{\partial y} + \overline{w}\frac{\partial \overline{S}}{\partial z} + \overline{u'\frac{\partial S'}{\partial x}} + \overline{v'\frac{\partial S'}{\partial y}} + \overline{w'\frac{\partial S'}{\partial z}} = \kappa_S \nabla^2 \overline{S} \tag{10.3}$$

10.3.1 Reynolds fluxes and eddy diffusivity

Using the continuity equation for the fluctuating velocity $(\nabla \cdot \vec{V}' = 0)$ we can rewrite the turbulent terms by adding $S'\nabla \cdot \vec{V}' = 0$ to the last three terms on the left-hand side of (10.3) which becomes

$$\frac{\partial\left(\overline{u'S'}\right)}{\partial x}+\frac{\partial\left(\overline{v'S'}\right)}{\partial y}+\frac{\partial\left(\overline{w'S'}\right)}{\partial z}.$$

The turbulent fluxes $\overline{u'S'}, \overline{v'S'}, \overline{w'S'}$ (also called the Reynolds fluxes) are related to the mean gradients in a similar fashion. The analogy gives

$$\overline{u'S'}=-K_{Sx}\frac{\partial\overline{S}}{\partial x};\ \overline{v'S'}=-K_{Sy}\frac{\partial\overline{S}}{\partial y};\ \overline{w'S'}=-K_{Sz}\frac{\partial\overline{S}}{\partial z} \qquad (10.4)$$

where K_{Sx}, K_{Sy} and K_{Sz} are kinematic ***eddy diffusivities*** (m^2s^{-1}).

The ranges of values for K_z and K_H are similar to those of A_z and A_H respectively because they are properties of the turbulent flow field.

$$\frac{dS}{dt}=K_H\left(\frac{\partial^2 S}{\partial x^2}+\frac{\partial^2 S}{\partial y^2}\right)+K_z\frac{\partial^2 S}{\partial z^2} \qquad (10.5)$$

Here S is now the average salinity. In the same manner, an equation for the average temperature is

$$\frac{dT}{dt}=K_H\left(\frac{\partial^2 T}{\partial x^2}+\frac{\partial^2 T}{\partial y^2}\right)+K_z\frac{\partial^2 T}{\partial z^2}+Q_T \qquad (10.6)$$

10.4 Thermoclines and the thermohaline circulation

Consider the steady-state case and ignore Q_T because we are interested in the main thermocline. The (10.6) becomes

$$\vec{V}_H\cdot\nabla_H T+w\frac{\partial T}{\partial z}=K_H\nabla_H^2 T+K_z\frac{\partial^2 T}{\partial z^2} \qquad (10.7)$$

Our knowledge of the deep circulation is not sufficient to allow us to drop any of the terms as being small. Stommel's model suggests that both advective terms are needed in a thermohaline circulation theory and at least the vertical diffusion term. Lateral diffusion may well be important too. One possible balances can produce a reasonable looking thermohaline structure.

The idea that vertical advection is balanced mainly by vertical diffusion with the other terms being fairly small has been considered a reasonable possibility for a long time.

$$\frac{\partial^2 T}{\partial z^2} = \frac{w}{K_z}\frac{\partial T}{\partial z} \tag{10.8}$$

Assuming w/K_z independent of z and $T = T_d$ for $z \ll -K_z/w$, then $T = T_d + (T_o - T_d)$ $\exp(w/K_z)$ where T_o is the temperature at $z = 0$ (taken to be at the bottom of the mixed layer). Adding a mixed layer on top and adjusting w/K_z one can produce a reasonable fit to observed vertical temperature profiles.

In the interior of the ocean we can use the geostrophic approximation and ignore the horizontal derivatives of α (Boussinesq approximation). We get

$$\beta v + f\left(\frac{\partial u}{\partial x} + \frac{\partial v}{\partial y}\right) = 0$$

Using the continuity equation gives

$$\beta v = f\left(\frac{\partial w}{\partial z}\right)$$

(1) To have north–south flow, w must vary with z. As argued earlier, we must have north–south flow in the interior to keep the relative vorticity small. In the interior, w is thought to be upward.

(2) It could increase from zero at great depth to a maximum in the thermocline and decrease to zero at the base of the mixed layer. For our solution to (10.8) we require w/K_z to be constant. Thus K_z would have to be a maximum in the thermocline too.

(3) This behavior for K_z seems contrary to the expectation that K_z would be lower in the region of strongest static stability. However, we do not know how K_z depends on height. Also, there are likely to be internal waves in the thermocline and breaking of these waves could lead to sufficient mixing to make K_z larger there.

(4) To keep things in perspective, remember that reasonable looking solutions have been found with $K_z = 0$. Clearly it is difficult to find satisfactory solutions even in the interior. To obtain a closed basin solution analytically lateral boundary layers would have to be added.

10.5 The mixed layer of the ocean

The top few tens of meters of the ocean are usually observed to be fairly well-mixed, i.e. the temperature and salinity are fairly uniform. Below this region there is a thermocline (and perhaps a halocline) and hence a pycnocline region. The top layer is the oceanic planetary boundary layer where vertical friction effects are important. It is also called the Ekman layer.

(1) Convergences and divergences in the layer lead to circulations in the deeper water (Ekman pumping effect).

(2) This is also the region of (biological) primary productivity. The depth of the layer and mixing up of nutrients from below will be important factors in determining the productivity.

(3) There are meteorological effects both for weather and climate. Solar radiation is first absorbed in the ocean's upper layer. A large part of the atmosphere's energy supply comes from heat exchange with this layer, mainly in the form of the latent heat of the water evaporated at the surface which is released when the water condenses higher in the atmosphere.

The conceptual model presented here comes from P. Niiler (1975). The temperature is assumed uniform within the mixed layer with a thin transition zone at the bottom where the temperature changes rapidly to the value in the thermocline below the mixed layer. The velocity is also assumed to be independent of depth throughout the bulk of the layer (U = constant in the mixed layer, U = 0 below thermocline).

With horizontal gradients assumed negligible, by continuity and w=0 at the surface there is no mean vertical velocity and all the non-linear terms involving the mean velocity vanish. The equations for the horizontal velocity are

$$\begin{aligned} \frac{\partial u}{\partial t} - f\, v &= -\frac{\partial \overline{(u'w')}}{\partial z} + F_x \\ \frac{\partial v}{\partial t} + f\, u &= -\frac{\partial \overline{(v'w')}}{\partial z} + F_y \end{aligned} \tag{10.10}$$

where F_x and F_y are damping terms added to make inertial oscillations die out.

In the temperature equation the advective terms vanish because of the assumption of no gradients and the consequence that w=0. The source term (solar radiation) is included in the specified surface heat flux since we are treating the layer as a whole and the temperature equation is

$$\frac{\partial T}{\partial t} + \frac{\partial \overline{(w'T')}}{\partial z} = 0 \tag{10.11}$$

Now in the layer u, v, T, F_x and F_y are independent of z (by assumption). Thus,

$\partial/\partial z$ of $\overline{u'w'}, \overline{v'w'},$ and $\overline{w'T'}$ must also be independent of z and the stresses and the heat flux are linear function of z.

There are four unknowns, u, v, T and the layer depth h, but only three equations. For closure, we add the equation for conservation of kinetic energy. Assume that a fraction of the energy input by the wind coming from the upper shear layer is available for mixing downward the thermocline. Part is used for overcoming stability in layer caused by heating (or cooling). This is an exchange of kinetic energy for potential energy.

Mixing is assumed to erode the thermocline only during the first half of a pendulum day after wind is turned on. By that time the initial rapid deepening has thickened the layer and reduced the velocity jump; the Richardson Number becomes too large and the static stability prevents mixing from this source.

The model works fairly well in the central Pacific during the heating season. However, if run for several years, mixed layer becomes too deep.

Conclusions: Note that there is no widely accepted model for the mixed layer.