Chapter 13

Deep Circulation in the Ocean

The direct forcing of the oceanic circulation by wind discussed in the last few chapters is limited mostly to the upper kilometer of the water column. Below a kilometer lies the vast water masses of the ocean extending to depths of 4–5 km. The water is everywhere cold, with a potential temperature less than 4° C. The water mass is formed when cold, dense water sinks from the surface to great depths at high latitudes. It spreads out from these regions to fill the ocean basins, and it eventually upwells through the thermocline over large areas of the ocean. It is this upwelling that drives the deep circulation. The vast deep ocean is usually referred to as the *abyss*, and the circulation as the *abyssal circulation*.

The most dense water at the sea surface, water that is dense enough to sink to the very bottom, is formed when frigid air blows across the ocean at high latitudes in the Atlantic. The wind cools and evaporates water. If the wind is sufficiently cold, sea ice forms, further increasing the salinity of the water because ice is fresher than sea water. At a few areas at high latitudes in the Atlantic in winter, the density of the water increases sufficiently for the water to sink deep into the ocean, reaching the ocean bottom. At other polar regions, cold, dense water is formed, but it is not quite salty enough to sink to the bottom. Because the Atlantic is so salty, the most dense water forms only in the Atlantic.

At mid and low latitudes, the density, even in winter, is sufficiently low that the water cannot sink more than a few hundred meters into the ocean. The only exception are some seas, such as the Mediterranean Sea, where evaporation is so great that the salinity of the water is sufficiently great for the water to sink to intermediate depths in the seas. If these seas are in communication with the open ocean, the waters formed in winter in the seas spreads out to intermediate depths in the ocean.

The sinking of cold dense water at high latitudes is due to temperature and salinity differences, and the sinking and spreading of cold water is known as the Thermohaline Circulation or the Meridional Overturning Circulation. In this chapter we will consider theories and observations of the circulation, and the influence of the circulation on climate.

13.1 Importance of the Thermohaline Circulation

The deep circulation which carries cold water from high latitudes in winter to lower latitudes throughout the world has very important consequences.

- 1. The contrast between the cold deep water and the warm surface waters determines the stratification of the oceans. Stratification strongly influences ocean dynamics.
- 2. The volume of deep water is far larger than the volume of surface water; and although currents in the deep ocean are relatively weak, they have transports comparable to the surface transports.
- 3. The deep circulation has important influences on Earth's heat budget and climate. The deep circulation varies over periods from decades to centuries to perhaps a thousand years, and this variability is thought to modulate climate over such time intervals. The ocean may be the primary cause of variability over times ranging from years to decades, and it may have helped modulate ice-age climate.

Two aspects of the deep circulation are especially important for understanding Earth's climate and its possible response to increased carbon dioxide CO_2 in the atmosphere: i) the ability of cold water to absorb CO_2 from the atmosphere, and ii) the ability of deep currents to modulate the heat transported from the tropics to high latitudes.

The Oceans as a Reservoir of Carbon Dioxide The oceans are the primary reservoir of readily available CO₂, an important greenhouse gas. The oceans contain 40,000 GtC of dissolved, particulate, and living forms of carbon. The land contains 2,200 GtC, and the atmosphere contains only 750 GtC. Thus the oceans hold 50 times more carbon than the air. Furthermore, the amount of new carbon put into the atmosphere since the industrial revolution, 150 GtC, is less than the amount of carbon cycled through the marine ecosystem in five years. (1 GtC = 1 gigaton of carbon = 10^{12} kilograms of carbon.) Carbonate rocks such as limestone, the shells of marine animals, and corals are other, much larger, reservoirs; but this carbon is locked up. It cannot be easily exchanged with carbon in other reservoirs.

More CO_2 dissolves in cold water than in warm water. Just imagine shaking and opening a hot can of $Coke^{\mathsf{TM}}$. The CO_2 from a hot can will spew out far faster than from a cold can. Thus the cold deep water in the ocean is the major reservoir of dissolved CO_2 in the ocean.

New CO_2 is released into the atmosphere when fossil fuels and trees are burned. Roughly half of the CO_2 released into the atmosphere quickly dissolves in the cold waters of the ocean which carry it into the abyss.

Forecasts of future climate change depend strongly on how much CO_2 is stored in the ocean and for how long. If little is stored, or if it is stored and

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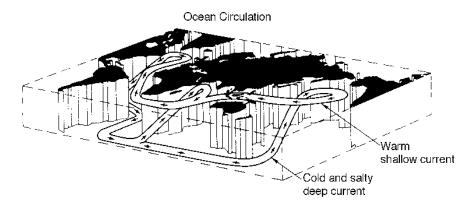


Figure 13.1 The oceanic conveyor belt carying heat northward into the north Atlantic. Note that this is a cartoon, and it does not accurately describe the ocean's circulation. (from Broecker and Peng, 1982).

later released into the atmosphere, the concentration in the atmosphere will change, modulating Earth's long-wave radiation balance. How much and how long CO_2 is stored in the ocean depends on the thermohaline circulation. The amount that dissolves depends on the temperature of the deep water, and the time the CO_2 is stored in the deep ocean depends on the rate at which deep water is replenished. Increased ventilation of deep layers, and warming of the deep layers could release large quantities of the gas to the atmosphere.

The storage of carbon in the ocean also depends on the dynamics of marine ecosystems, upwelling, and the amount of dead plants and animals stored in sediments; but we won't consider these processes.

Oceanic Transport of Heat The oceans carry about half the heat from the tropics to high latitudes required to maintain Earth's temperature. Heat carried by the Gulf Stream and the north Atlantic drift warms Europe. Norway, at 60° N is far warmer than southern Greenland or northern Labrador at the same latitude; and palm trees grow on the west coast of Ireland, but not in Newfoundland which is further south.

Wally Broecker (1982), who works at Lamont-Doherty Geophysical Observatory of Columbia University, calls the oceanic component of the heat-transport system the *Global Conveyor Belt*. The basic idea is that the Gulf Stream carries heat to the far north Atlantic. There the surface waters release heat and water to the atmosphere and become sufficiently dense that they sink to the bottom in the in the Norwegian Sea and in the Greenland Sea. The deep water later upwells in other regions and in other oceans, and eventually makes its way back to the Gulf Stream and the north Atlantic.

A simple picture of the conveyor belt (Figure 13.1) shows the important elements of the transports. Nevertheless, it is a cartoon, a much simplified illustration of reality. For example, the picture ignores the large sources of bottom water offshore of Antarctica in the Weddel and Ross Seas. We can make a crude estimate of the importance of the conveyor belt circulation from a simple calculation based on what we know about waters in the Atlantic compiled by Bill Schmitz (1996) in his wonderful summary of his life's work. The Gulf Stream carries 40 Sv of 18°C water northward. Of this, 14 Sv return southward in the deep western boundary current at a temperature of 2°C. The flow carried by the conveyor belt must therefore lose 0.9 petawatts (1 petawatt = 10^{15} watt) in the north Atlantic north of 24°N. Although the calculation is very crude, it is remarkably close to the value of 1.2 ± 0.2 petawatts estimated much more carefully by Rintoul and Wunsch (1991).

Note that if the water remained on the surface and returned as an eastern boundary current, it would be far warmer than the deep current when it returned southward. Hence, the heat transport would be much reduced.

So much heat is transported northward in the north Atlantic that heat transport in the Atlantic is entirely northward, even in the southern hemisphere (figure 5.12). Much of the solar heat absorbed by the tropical Atlantic is shipped north to warm Europe and the northern hemisphere. Imagine then what might happen if the supply of heat is shut off. We will get back to that topic in the next section.

The production of bottom water is modulated by slow changes of surface salinity in the north Atlantic. It is also modulated by the rate of upwelling due to mixing in other oceanic areas. First, let's look at the influence of salinity.

More saline surface waters form denser water in winter than less saline water. At first you may think that temperature is also important, but at high latitudes water in all oceans becomes cold enough to freeze, so all oceans produce $-2^{\circ}C$ water at the surface. Of this, only the most salty will sink, and the saltiest water is in the Atlantic.

The production of bottom water is remarkably sensitive to small changes in salinity. Rahmstorf (1995), using a numerical model of the meridional-overturning circulation, showed that a ± 0.1 Sv variation of the flow of fresh water into the north Atlantic can switch on or off the deep circulation of 14 Sv. If the deep-water production is shut off during times of low salinity, the 1 petawatt of heat may also be shut off.

I write may be shut off because the ocean is a very complex system. We don't know if other processes will increase heat transport if the deep circulation is disturbed. For example, the circulation at intermediate depths may increase when deep circulation is reduced.

The production of bottom water is also remarkably sensitive to small changes in mixing in the deep ocean. Munk and Wunsch (1998) calculate that 2.1 TW (terawatts = 10^{12} watts) are required to drive the deep circulation; and that this small source of mechanical mixing drives a poleward heat flux of 2000 TW. Most of the energy for mixing comes from dissipation of tidal currents, which depend on the distribution of the continents. Thus during the last ice age, when sea level was much lower, tides, tidal currents, tidal dissipation, and deep circulation would all differ from present values. Role of the Ocean in Ice-Age Climate Fluctuations What might happen when the production of deep water in the Atlantic is shut off? Information contained in the Greenland and Antarctic ice sheets and in north Atlantic sediments provide important clues.

Two ice cores through the Greenland ice sheet and three through the Antarctic sheet provide a continuous record of atmospheric conditions over Greenland and Antarctica extending back more than a hundred-thousand years before the present. Annual layers in the core are counted to get age. Deeper in the core, where annual layers are hard to see, age is calculated from depth. The oxygenisotope ratios in the ice give temperatures over parts of the northern hemisphere, bubbles in the ice give atmospheric CO_2 concentration, and chemical composition and particle concentration give information about volcanic eruptions and windiness of the atmosphere.

Cores through deep-sea sediments in the north Atlantic made by the Ocean Drilling Program give information about sea-surface above the core, the production of north Atlantic deep water, and the production of icebergs.

- 1. The oxygen-isotope record in the ice cores show remarkable temperature variability over the past 100,000 years. Many times during the last ice age, temperatures near Greenland warmed rapidly over periods of 1–100 years, followed by gradual cooling over longer periods (Dansgaard et al, 1993). For example, $\approx 11,500$ years ago, temperatures over Greenland warmed by $\approx 8^{\circ}$ C in 40 years in three steps, each spanning 5 years (Alley, 2000). Such abrupt warming is called a Dansgaard/Oeschger event. Other studies have shown that much of the northern hemisphere warmed and cooled in phase with temperatures calculated from the ice core.
- 2. Hartmut Heinrich and colleagues (Bond et al. 1992), studying the sediments in the north Atlantic noticed periods when coarse material was deposited on the bottom in mid ocean. Only icebergs can carry such material out to sea, and the find indicated times when large numbers of icebergs were released into the north Atlantic. These are now called Heinrich events.
- 3. Heinrich events seem to preceed the largest Dansgaard/Oeschger events. The Dansgaard/Oeschger–Heinrich tandem events seem to be related to warming events seen in Antarctic ice cores. Temeratures changes in the two hemispheres are out of phase. When Greenland warms, Antarctica cools.
- 4. The correlation of Greenland temperature with iceberg production is related to the meridional overturning circulation. When icebergs melted, the surge of fresh water increased the stability of the water column shutting off the production of North Atlantic Deep Water. The shut-off of deep-water formation greatly reduced the transport of warm water in the north Atlantic, producing very cold northern hemisphere climate (Figure 13.2). The melting of the ice pushed the polar front, the boundary between cold and warm water in the north Atlantic further south than its

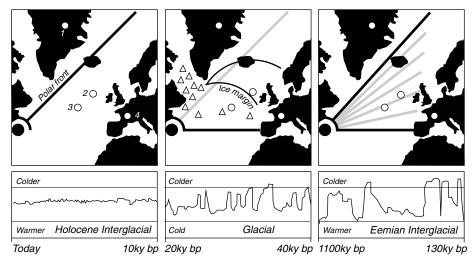


Figure 13.2 Periodic surges of icebergs during the last ice age appear to have modulated temperatures of the northern hemisphere by lowering the salinity of the far north Atlantic and reducing the meridional overturning circulation. Data from cores through the Greenland ice sheet (1), deep-sea sediments (2,3), and alpine-lake sediments (4) indicate that: Left: During recent times the circulation has been stable, and the polar front which separates warm and cold water masses has allowed warm water to penetrate beyond Norway. Center: During the last ice age, periodic surges of icebergs reduced salinity and reduced the meridional overturning circulation, causing the polar front to move southward and keeping warm water south of Spain. Right: Similar fluctuations during the last intergalcial appear to have caused rapid, large changes in climate. The Bottom plot is a rough indication of temperature in the region, but the scales are not the same (from Zahn, 1994).

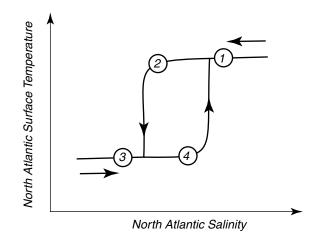


Figure 13.3 The meridional-overturning circulationis part of a non-linear system. The circulation has two stable states near 2 and 4. The switching of north Atlantic from a warm, salty regime to a cold, fresh regime and back has hysterisis. This means that as the warm salty ocean in an initial state 1 freshens, and becomes more fresh than 2 it quickly switches to a cold, fresh state 3. When the area again becomes salty, it must move past state 4 before it can switch back to 1.

present position. The location of the front, and the time it was at different positions can be determined from analysis of bottom sediments.

- 5. When the meridional overturning circulation shuts down, heat normally carried from the south Atlantic to the north Atlantic becomes available to warm the southern hemisphere. This explains the Antarctic warming.
- 6. The switching on and off of the meridional overturning circulation has large hysterisis (Figure 13.3). The circulation has two stable states. The first is the present circulation. In the second, deep water is produced mostly near Antarctica, and upwelling occurs in the far north Pacific (as it does today) and in the far north Atlantic. Once the circulation is shut off, th system switches to the second stable state. The return to normal salinity does not cause the circulation to turn on. Surface waters must become saltier than average for the first state to return (Rahmstorf, 1995).
- 7. A weakened version of this process with a period of about 1000 years may be modulating present-day climate in the north Atlantic, and it may have been responsible for the Little Ice Age from 1100 to 1800.

The simple picture we have painted of the variability in salinity, air temperature, and deep-water formation is not yet well understood. For example, we don't know what causes the ice sheets to surge. Surges may result from warmer temperatures caused by increased water vapor from the tropics (a greenhouse gas) or from an internal instability of a large ice sheet. Nor do we know exactly how the oceanic circulation responds to changes in the deep circulation or surface moisture fluxes. Recent work by Wang, Stone and Marotzke (1999), who used a numerical model to simulate the climate system, shows that the meridional overturning circulation is modulated by moisture fluxes in the *southern* hemisphere.

13.2 Theory for the Thermohaline Circulation

Stommel, Arons, and Feller in a series of papers from 1958 to 1960 laid the foundation for our present understanding of the abyssal circulation (Stommel 1958; Stommel, Arons, and Faller, 1958; Stommel and Arons, 1960). The papers reported simplified theories of the circulation that differed so greatly from what was expected that Stommel and Arons devised laboratory experiments with rotating fluids to confirmed their theory. The theory for the deep circulation has been further discussed by Marotzke (2000) and Munk and Wunsch (1998).

The theory Stommel, Arons, Feller theory is based on three fundamental ideas:

- 1. The source of cold, deep water is sinking at a few locations at high latitudes in the Atlantic, notably in the Irminger and Greenland Seas in the north and the Weddel Sea in the south.
- 2. The thermocline is remarkably sharp everywhere in mid to low latitudes, and the shape of the profile does not change from year to year. Because turbulence in the thermocline transports heat downward, the thermocline should become much weaker after many years. Since this does not happen,

upwelling and mixing in the thermocline must carry cold water upward at a rate that balances the downward flux of heat.

3. The abyssal circulation is strictly geostrophic in the interior of the ocean, and therefore potential vorticity is conserved.

Notice that the deep circulation is not driven by convection. Munk and Wunsch (1998) point out that deep convection by itself leads to a deep, stagnant, pool of cold water. In this case, the thermohaline circulation is confined to the upper layers of the ocean. Mixing or upwelling is required to pump cold water upward through the thermocline and drive the meriodional overturning circulation.

Notice also that convection and sinking are not the same, and they do not occur in the same place (Marotzke and Scott, 1999). Convection ocurs in small regions a few kilometers on a side. Sinking, driven by Ekman pumping and geostrophic currents, can occur over far larger areas. In this chapter, we are discussing mostly sinking of water.

To describe the simplest aspects of the flow, we begin with the Sverdrup equation applied to a bottom current of thickness H in an ocean of constant depth:

$$\beta v = f \frac{\partial w}{\partial z} \tag{13.1}$$

where $f = 2\Omega \sin \varphi$, $\beta = (2\Omega \cos \varphi)/R$, Ω is Earth's rotation rate, R Earth's radius, and φ is latitude. Integrating (13.1) from the bottom of the ocean to the top of the abyssal circulation gives:

$$V = \int_{0}^{H} v \, dz = \int_{0}^{H} \frac{f}{\beta} \, \frac{\partial w}{\partial z} \, dz$$
$$V = \frac{R \tan \varphi}{H} W_{0}$$
(13.2)

where V is the vertical integral of the northward velocity, and W_0 is the velocity at the base of the thermocline. Because the vertical velocity is everywhere positive as required to balance the downward diffusion of heat, then V must be everywhere toward the poles. This is the abyssal flow in the interior of the ocean sketched by Stommel in Figure 13.4. The U component of the flow is calculated from V and w using the continuity equation.

To connect the streamlines of the flow in the west, Stommel added a deep western boundary current. The strength of the western boundary current depends on the volume of water S produced at the source regions. Stommel and Arons calculated the flow for a simplified ocean bounded by the Equator and two meridians (a pie shaped ocean). First they placed the source S_0 near the pole to approximate the flow in the north Atlantic. If the volume of water sinking at the source equals the volume of water upwelled in the basin, and if the upwelled velocity is constant everywhere, then the transport T_w in the western boundary current is:

$$T_w = -2\,S_0\sin\varphi\tag{13.3}$$

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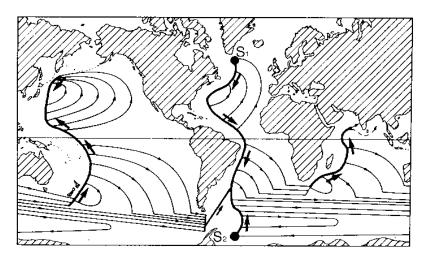


Figure 13.4 Sketch of the deep circulation resulting from deep convection in the Atlantic (dark circles) and upwelling through the thermocline elsewhere (After Stommel, 1958).

The transport in the western boundary current at the poles is twice the volume of the source, and the transport diminishes to zero at the Equator (Stommel and Arons, 1960a: eq, 7.3.15; see also Pedlosky, 1996: §7.3). The flow driven by the upwelling water adds a recirculation equal to the source. If S_0 exceeds the volume of water upwelled in the basin, then the western boundary current carries water across the Equator. This gives the western boundary current sketched in the north Atlantic in Figure 13.4.

Next, Stommel and Arons calculated the transport in a western boundary current in a basin with no source. The transport is:

$$T_w = S\left[1 - 2\sin\varphi\right] \tag{13.4}$$

where S is the transport across the Equator from the other hemisphere. In this basin Stommel notes:

A current of recirculated water equal to the source strength starts at the pole and flows toward the source ... [and] gradually diminishes to zero at $\varphi = 30^{\circ}$ north latitude. A northward current of equal strength starts at the equatorial source and also diminishes to zero at 30° north latitude.

This gives the western boundary current as sketched in the north Pacific in Figure 13.4.

Note that the Stommel-Aron theory assumes a flat bottom. The mid-ocean ridge system divides the deep ocean into a series of basins connected by sills through which the water flows from one basin to the next. As a result, the flow in the deep ocean is not as simple as that sketched by Stommel. Boundary current flow along the edges of the basins, and flow in the eastern basins in the Atlantic comes through the mid-Atlantic ridge from the western basics. Figure 13.5 shows how ridges control the flow in the Indian Ocean. Nevertheless, the basic flow is remarkably like that suggested by Stommel.

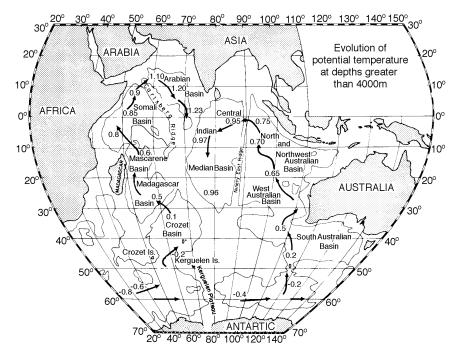


Figure 13.5 Sketch of the deep circulation in the Indian Ocean inferred from the temperature, given in °C. Note that the flow is constrained by the deep mid-ocean ridge system (After Tchernia, 1980).

Finally, the Stommel-Arons theory gives some values for time required for the water to move from the source regions to the base of the thermocline in various basins. The time required varies from a few hundred years for basins near the sources to nearly a thousand years for the north Pacific, which is fartherest from the sources.

Some Comments on the Theory for the Deep Circulation Our understanding of the deep circulation is still evolving.

- 1. The deep circulation is driven by mixing in the ocean, not by deep convection near the poles. Marotzke and Scott (1999) points out that the two processes are very different. Convection reduces the potential energy of the water column, and it is self powered. Mixing in a stratified fluid increases the potential energy, and it must be driven by an external process.
- 2. Numerical models of the deep circulation show that the meridional overturning circulation is very sensitive to the assumed vertical diffusivity coefficient in the thermocline (Gargett and Holloway, 1992).
- 3. Numerical calculations by Marotzke and Scott (1999) indicate that the transport is not limited by the rate of deep convection, but it is sensitive to the assumed vertical mixing coefficient, especially near side boundaries.
- 4. Where is cold water mixed upward? Is it in the thermocline or at the ocean's boundaries? Recent measurements of vertical mixing (§8.5) sug-

gest mixing is concentrated above seamounts and mid-ocean ridges, and along strong currents such as the Gulf Stream.

5. Because the meridional overturning circulation is pulled by mixing and not pushed by deep convection, the transport of heat into the north Atlantic may not be as sensitive to surface salinity as described above.

13.3 Observations of the Deep Circulation

The abyssal circulation is less well known than the upper-ocean circulation. Direct observations from moored current meters or deep-drifting floats were difficult to make until recently, and there are few long-term direct measurements of current. In addition, the measurements do not produce a stable mean value for the deep currents. For example, if the deep circulation takes roughly 1,000 years to transport water from the north Atlantic to the Antarctic Circumpolar Current and then to the north Pacific, the mean flow is about 1 mm/s. Observing this small mean flow in the presence of typical deep currents having variable velocities of up to 10 cm/s or greater, is very difficult.

Most of our knowledge of the deep circulation is inferred from measured distribution of temperature, salinity, oxygen, silicate, tritium, fluorocarbons and other tracers. These measurements are much more stable than direct current measurements, and observations made decades apart can be used to trace the circulation. Tomczak (1999) carefully describes how the techniques can be made quantitative and how they can be applied in practice.

Water Masses The concept of water masses originates in meteorology. Vilhelm Bjerknes, a Norwegian meteorologist, first described the cold air masses that form in the polar regions. He showed how they move southward, where they collide with warm air masses at places he called fronts, just as masses of troops collide at fronts in war (Friedman, 1989). In a similar way, water masses are formed in different regions of the ocean, and the water masses are often separated by fronts. Note, however, that strong winds are associated with fronts in the atmosphere because of the large difference in density and temperature on either side of the front. Fronts in the ocean sometimes have little contrast in density, and these fronts have only weak currents.

Tomczak (1999) defines a water mass as a

body of water with a common formation history, having its origin in a physical region of the ocean. Just as air masses in the atmosphere, water masses are physical entities with a measureable volume and therefore occupy a finite volume in the ocean. In their formation region they have exclusive occupation of a particular part of the ocean; elsewhere they share the ocean with other water masses with which they mix. The total volume of a water mass is given by the sum of all its elements regardless of their location.

Plots of salinity as a function of temperature, called T-S plots, are used to delineate water masses and their geographical distribution, to describe mixing among water masses, and to infer motion of water in the deep ocean. Here's

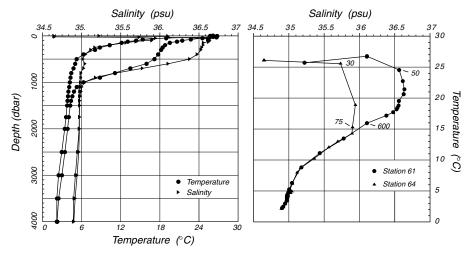


Figure 13.6 Temperature and salinity measured at hydrographic stations on either side of the Gulf Stream. Data are from Tables 10.2 and 10.4. Left: Temperature and salinity plotted as a function of depth. **Right:** The same data, but salinity is plotted as a function of temperature in a T-S plot. Notice that temperature and salinity are uniquely related below the mixed layer. A few depths are noted next to data points.

why the plots are so useful: water properties, such as temperature and salinity, are formed only when the water is at the surface or in the mixed layer. Heating, cooling, rain, and evaporation all contribute. Once the water sinks below the mixed layer temperature and salinity can change only by mixing with adjacent water masses. Thus water from a particular region has a particular temperature associated with a particular salinity, and the relationship changes little as the water moves through the deep ocean.

Thus temperature and salinity are not independent variables. For example, the temperature and salinity of the water at different depths below the Gulf Stream are uniquely related (Figure 13.6, right), indicating they came from the same source region, even though they do not appear related if temperature and salinity are plotted independently as a function of depth (Figure 13.6, left).

Temperature and salinity are *conservative properties* because there are no sources or sinks of heat and salt in the interior of the ocean. Other properties, such as oxygen are non-conservative. For example, oxygen content may change slowly due to oxidation of organic material and respiration by animals.

Each point in the T-S plot is called a *water type*. This is a mathematical ideal. Some water masses may be very homogeneous and they are almost points on the plot. Other water masses are less homogeneous, and they occupy regions on the plot.

The mixing of two water types leads to a straight line on a T-S diagram (Figure 13.7). Because the lines of constant density on a T-S plot are curved, mixing increases the density of the water. This is called *densification* (Figure 13.8).

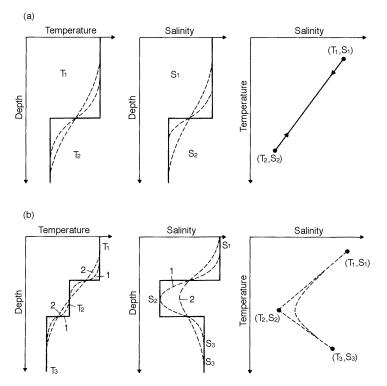


Figure 13.7 **Upper:** Mixing of two water masses produces a line on a T-S plot. Lower: Mixing among three water masses produces intersecting lines on a T-S plot, and the apex at the intersection is rounded by further mixing. From Tolmazin (1985).

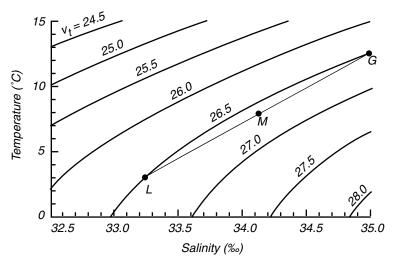


Figure 13.8 Mixing of two water types of the same density produces water that is denser than either water type. From Tolmazin (1985).

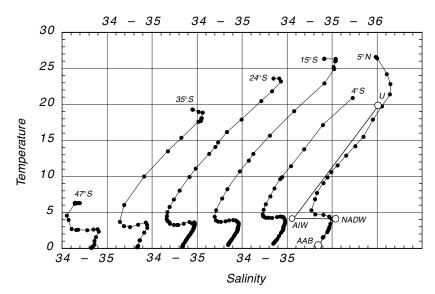


Figure 13.9 T-S plot of data collected at various latitudes in the western basins of the south Atlantic. Lines drawn through data from 5°N, showing possible mixing between water masses: NADW-North Atlantic Deep Water, AIW-Antarctic Intermediate Water, AAB Antarctic Bottom Water, U Subtropical Lower Water.

Water Masses and the Deep Circulation Let's use these ideas of water masses and mixing to study the deep circulation. We start in the south Atlantic because it has very clearly defined water masses. A T-S plot calculated from hydrographic data collected in the south Atlantic (Figure 13.9) shows three important water masses listed in order of decreasing depth (Table 13.1): Antarctic Bottom Water AAB, North Atlantic Deep Water NADW, and Antarctic Intermediate Water AIW. All are deeper than one kilometer. The mixing among three water masses shows the characteristic rounded apexes shown in the idealized case shown in Figure 13.7.

The plot indicates that the same water masses can be found throughout the western basins in the south Atlantic. Now let's use a cross section of salinity to trace the movement of the water masses using the core method.

Core Method The slow variation from place to place in the ocean of a tracer such as salinity can be used to determine the source of the waters masses such as those in Figure 13.8. This is called the *core method*. The method may also be used to track the slow movement of the water mass. Note, however, that a slow drift of the water and horizontal mixing both produce the same observed properties in the plot, and they cannot be separated by the core method.

A core is a layer of water with extreme value (in the mathematical sense) of salinity or other property as a function of depth. An extreme value is a local maximum or minimum of the quantity as a function of depth. The method assumes that the flow is along the core. Water in the core mixes with the water masses above and below the core and it gradually loses its identity. Furthermore, the flow tends to be along surfaces of constant potential density.

			Temp. (°C)	Salinity (psu)
Antarctic water	Antarctic Intermediate Water Antarctic Bottom Water	AIW ABW	$3.3 \\ 0.4$	$34.15 \\ 34.67$
North Atlantic water	North Atlantic Deep Water North Atlantic Bottom Water	NADW NABW	$4.0 \\ 2.5$	$35.00 \\ 34.90$
Thermocline water	Subtropical Lower Water	U	18.0	35.94

Table 13.1 Water Masses of the south Atlantic between $33^{\circ}S$ and $11^{\circ}N$

From Defant (1961: Table 82)

Let's apply the method to the data from the south Atlantic to find the source of the water masses. As you might expect, this will explain their names.

We start with a north-south cross section of salinity in the western basins of the Atlantic (Figure 13.10). It we locate the maxima and minima of salinity as a fucntion of depth at different latitudes, we can see two clearly defined cores. The upper low-slainity core starts near 55°S and it extends northward at depths near 1000 m. This water originates at the Antarctic Polar Front zone. This is the Antarctic Intermediate Water. Below this water mass is a core of salty water originating in the north Atlantic. This is the North Atlantic Deep Water. Below this is the most dense water, the Antarctic Bottom Water. It originates in winter when cold, dense, saline water forms in the Weddel Sea and other shallow seas around Antarctica. The water sinks along the continental slope and mixes with Circumpolar Deep Water. It then fills the deep basins of the south Pacific, Atlantic, and Indian Oceans.

The Circumpolar Deep Water is mostly North Atlantic Deep Water that has been carried around Antarctica. As it is carried along, it mixes with deep waters of the Indian and Pacific Oceans to form the circumpolar water.

The flow is probably not along the arrows shown in Figure 13.10. The

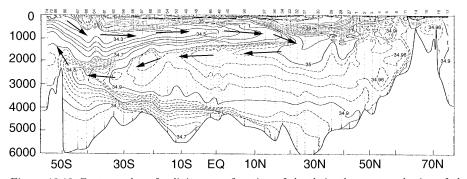


Figure 13.10 Contour plot of salinity as a function of depth in the western basins of the Atlantic from the Arctic Ocean to Antarctica. The plot clearly shows extensive cores, one at depths near 1000 m extending from 50° S to 20° N, the other at is at depths near 2000 m extending from 20° N to 50° S. The upper is the Antarctic Intermediate Water, the lower is the North Atlantic Deep Water. The arrows mark the assumed direction of the flow in the cores. The Antarctic Bottom Water fills the deepest levels from 50° S to 30° N. See also Figures 10.16 and 6.11. From Lynn and Ried (1968).

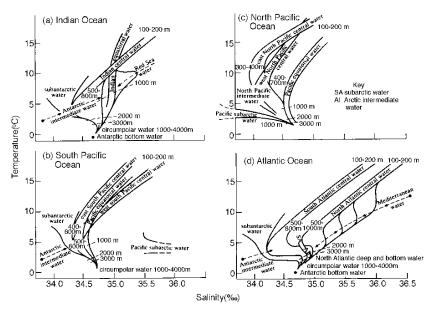


Figure 13.11 T-S plots of water in the various ocean basins. From Tolmazin (1985).

distribution of properties in the abyss can be explained by a combination of slow flow in the direction of the arrows plus horizontal mixing along surfaces of constant potential density with some weak vertical mixing. The vertical mixing probably occurs at the places where the density surface reaches the sea bottom at a lateral boundary such as seamounts, mid-ocean ridges, and along the western boundary. Flow in a plane perpendicular to that of the figure may be at least as strong as the flow in the plane of the figure shown by the arrows.

The core method can be applied only to a tracer that does not influence density. Hence temperature is usually a poor choice. If the tracer controls density, then flow will be around the core according to ideas of geostrophy, not along core as assumed by the core method.

The core method works especially well in the south Atlantic with its clearly defined water masses. In other ocean basins, the T-S relationship is more complicated. The abyssal waters in the other basins are a complex mixture of waters coming from different areas in the ocean (Figure 13.11). For example, warm, salty water from the Mediterranean Sea enters the north Atlantic and spreads out at intermediate depths displacing intermediate water from Antarctica in the north Atlantic, adding additional complexity to the flow as seen in the lower right part of the figure.

Other Tracers I have illustrated the core method using salinity as a tracer, but many other tracers are used. An ideal tracer is easy to measure even when its concentration is very small; it is conserved, which means that only mixing changes its concentration; it does not influence the density of the water; it exists in the water mass we wish to trace, but not in other adjacent water masses; and it does not influence marine organisms (we don't want to release toxic tracers).

Various tracers meet these criteria to a greater or lesser extent, and they are used to follow the deep and intermediate water in the ocean. Here are some of the most widely used tracers.

- 1. Salinity is conserved, and it influences density much less than temperature.
- 2. Oxygen is only partly conserved. Its concentration is reduced by the respiration of marine animals and the oxidation of marine products.
- 3. Silicates are used by some marine organisms. They are conserved at depths below the sunlit zone.
- 4. Phosphates are used by all organisms, but they can provide additional information.
- 5. ³He is conserved, but there are few sources, mostly at deep-sea volcanic areas and hot springs.
- 6. ³H (tritium) was produced by atomic bomb tests in the atmosphere in the 1950s. It enters the ocean through the mixed layer, and it is useful for tracing the formation of deep water. It decays with a half life of 12.3 y and it is slowly disappearing from the ocean. Figure 10.16 shows the slow advection or perhaps mixing of the tracer into the deep north Atlantic. Note that after 25 years little tritium is found south of 30° N. This implies a mean velocity of less than a mm/s.
- 7. Fluorocarbons (Freons used in air conditioning) have been recently injected into atmosphere. They can be measured with very great sensitivity; and they are being used for tracing the sources of deep water.
- 8. Sulpher hexafluoride SF_6 can be injected into sea water, and the concentration can be measured with great sensitivity for many months.

Each tracer has its usefulness, and each provides additional information about the flow.

13.4 Antarctic Circumpolar Current

The Antarctic Circumpolar Current is an important feature of the ocean's deep circulation because it transports deep and intermediate water between the Atlantic, Indian, and Pacific Ocean, and because it contributes to the deep circualtion in all basins. Because it is so important for understanding the deep circulation in all oceans, let's look at what is known about this current.

A plot of density across a line of constant longitude in the Drake Passage (Figure 13.12) shows three fronts. They are, from north to south: 1) the Subantarctic Front, 2) the Polar Front, and 3) the Southern ACC Front. Each front is continuous around Antarctica (Figure 13.13). The plot also shows that the constant-density surfaces slope at all depths, which indicates that the currents extend to the bottom.

Typical current speeds are around 10 cm/s with speeds of up to 50 cm/s near some fronts. Although the currents are slow, they transport much more water than western boundary currents because the flow is deep and wide. Whitworth

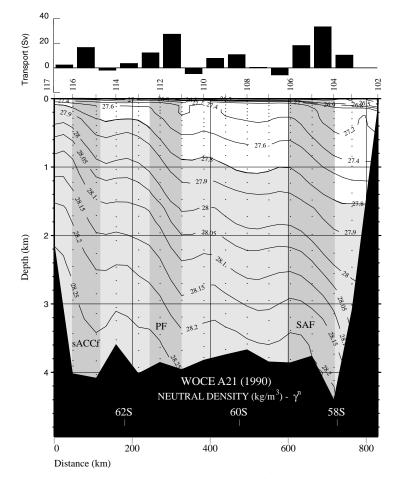


Figure 13.12 Cross section of neutral density across the Antarctic Circumpolar Current in the Drake Passage from the World Ocean Circulation Experiment section A21 in 1990. The current has three streams associated with the three fronts (dard shading): sF = Southern ACC Front, PF = Polar Front, and sAF = Subantarctic Front. Hydrographic station numbers are given at the top, and transports are relative to 3,000 dbar. Circumpolar deep water is indicated by light shading. From Orsi (2000)

and Peterson (1985) calculated transport through the Drake Passage using several years of data from an array of 91 current meters on 24 moorings spaced approximately 50 km apart along a line spanning the passage. They also used measurements of bottom pressure measured by gauges on either side of the passage. They found that the average transport through the Drake Passage was 125 ± 11 Sv, and that the transport varied from 95 Sv to 158 Sv. The maximum transport tended to occur in late winter and early spring (Figure 13.14).

Because the antarctic currents extend all the way to the bottom, they are influenced by topographic steering. As the current crosses ridges such as the Kerguelen Plateau, the Pacific-Antarctic Ridge, the Drake Passage, it is deflected by the ridges.

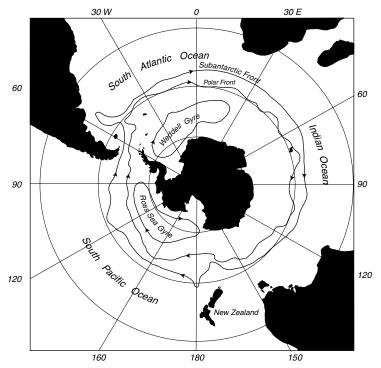


Figure 13.13 Distribution of the Subantarctic and Polar Fronts and associated currents in the Antarctic. From Whitworth (1988)

The core of the current is composed of circumpolar deep water, a mixture of deep water from all oceans. The upper branch of the current contains oxygenpoor water from all oceans. The lower (deeper) branch contains a core of highsalinity water from the Atlantic, including contributions from the north Atlantic deep water mixed with salty Mediterranean Sea water. As the different water masses circulate around Antarctica they mix with other water masses with similar density. In a sense, the current is a giant 'mix-master' taking deep water from each ocean, mixing it with deep water from other oceans, and then redistributing it back to each ocean.

The coldest, saltiest water in the ocean is produced on the continental shelf around Antarctica in winter, mostly from the shallow Weddel and Ross seas. The cold salty water drains from the shelves, entrains some deep water, and spreads out along the sea floor. Eventually, 8–10 Sv of bottom water are formed (Orsi, Johnson, and Bullister, 1999). This dense water then seeps into all the ocean basins. By definition, this water is too dense to cross pass through thee Drake Passage, so it is not circumpolar water.

The Antarctic currents are wind driven. Strong west winds with maximum speed near 50°S drive the currents (see Figure 4.2), and the north-south gradient of wind speed produces convergence and divergence of Ekman transports. Divergence south of the zone of maximum wind speed, south of 50°S leads to upwelling of the circumpolar deep water. Convergence north of the zone

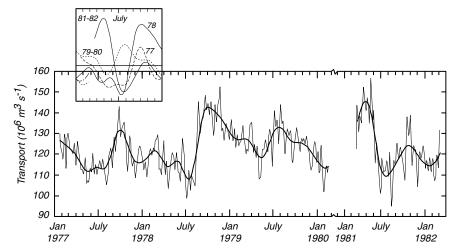


Figure 13.14 Variability of the transport in the Antarctic Circumpolar Current as measured by an array of current meters deployed across the Drake Passage. The heavier line is smoothed, time-averaged transport. From Whitworth (1988)

of maximum winds leads to downwelling of the Antarctic intermediate water. The surface water is relatively fresh but cold, and when they sink they define characteristics of the Antarctic intermediate water.

Because wind constantly transfers momentum to the circumpolar current, causing it to accelerate, the acceleration must be balanced by some type of drag, and we are led to ask: What keeps the flow from acelerating to very high speeds? The simple answer, from Munk and Palmen (1951), is form drag. Form drag is due to the current crossing subsea ridges, especially at the Drake Passage. Form drag is also the drag of the wind on a fast moving car. In both cases, the flow is diverted, by the ridge or by your car, creating a low pressure zone downstream of the ridge or down wind of the car. The low pressure zone transfers momentum into the solid earth, slowing down the current.

13.5 Important Concepts

- 1. The deep circulation of the ocean is very important because it determines the vertical stratification of the oceans and because it modulates climate.
- 2. The cold, deep water in the ocean absorbs CO_2 from the atmosphere, therefore temporarily reducing atmospheric CO_2 . Eventually, however, most of the CO_2 must be released back to the ocean. (Some is used by plants, some is used to make sea shells).
- 3. The production of deep bottom waters in the north Atlantic causes a transport of one petawatt of heat into the northern hemisphere which warms Europe.
- 4. Variability of deep water formation in the north Atlantic has been tied to large fluctuations of northern hemisphere temperature during the last ice ages.

- 5. The theory for the deep circulation was worked out by Stommel and Aarons in a series of papers published from 1958 to 1960. They showed that vertical velocities are needed nearly everywhere in the ocean to maintain the thermocline, and the vertical velocity drives the deep circulation.
- 6. The deep circulation is driven by vertical mixing, which is largest above mid-ocean ridges, near seamounts, and in strong boundary currents.
- 7. The deep circulation is too weak to measure directly. It is inferred from observations of water masses defined by their temperature and salinities and from observation of tracers.
- 8. The Antarctic Circumpolar Current mixes deep water from all oceans and redistributes it back to each ocean. The current is deep and slow with a transport of 125 Sv.