Appendix A Ocean Circulation

(10/15/01)

The chemistry and biology of the ocean are superimposed on the ocean's circulation, thus it is important to review briefly the forces driving this circulation and give some estimates of the transport rates. There are many reasons why it is important to understand the basics of the circulation. Four examples are given as an illustration.

(1) Poleward-flowing, warm, surface, western boundary currents such as the Gulf Stream and the Kuroshiro have a profound effect on the seasurface temperature (SST) and the climate of land areas bordering the oceans. For example, the Gulf Stream transports approximately 3.2 peta Watts (peta = 10^{15}) of heat to the North Atlantic (Hartmann, 1994), moderating the climate of northern Europe.

(2) The El-Nino Southern Oscillation (ENSO) phenomenon is an interannual perturbation of the climate system characterized by aperiodic weakening of the tradewinds and warming of the surface water in the central and eastern equatorial Pacific Ocean. The impacts of ENSO are felt worldwide through disruption of atmospheric circulation and weather patterns (McPhaden, 1993; Wallace et al, 1998)

(3) Atmospheric testing of nuclear bombs resulted in the contamination of the surface of the ocean with various isotopes including ^{14}C, ^{3}H, ^{90}Sr, ^{239}Pu, and ^{240}Pu that have different chemistries and cource functions. These isotopes are slowly being mixed through the ocean and can be used as radioactive dyes and clocks (e.g. Broecker and Peng, 1982).

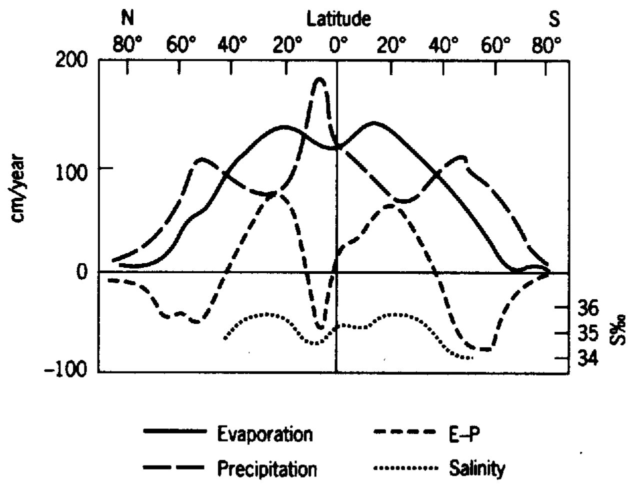
(4) The atmospheric CO_2 concentration has been increasing since the beginning of the industrial age, but the increase (~3.2 Gt C yr^{-1}) is less than the sum of anthropogenic emissions and deforestation (~ 7.0 Gt C yr^{-1}) (Siegenthaler and Sarmiento, 1993). Some of the CO_2 has gone into the ocean (~2 Gt C yr^{-1}). All CO_2 taken up by the ocean is done so by the process of gas-exchange. Some of the excess CO_2 has been transported into the intermediate and deep water by the subduction of water masses. Circulation replenishes the surface with water undersaturated with respect to the anthropogenically perturbed CO_2 levels.

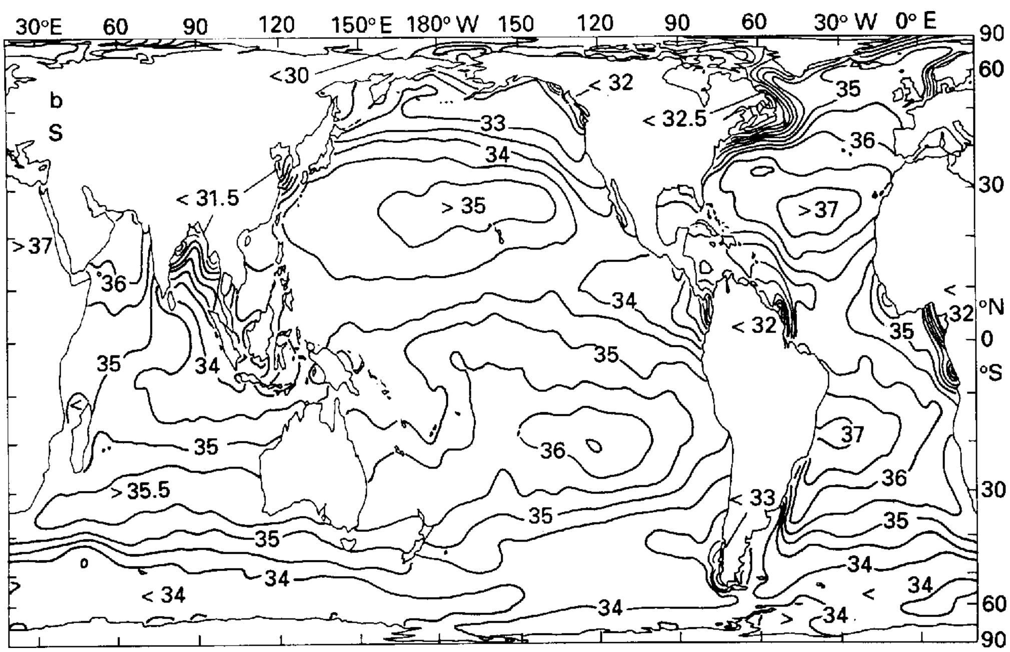
In this Appendix we give a brief overview of what controls the density of seawater and the vertical density stratification of the ocean. Surface currents, abyssal circulation, and thermocline circulation will be considered individually. For more thorough discussion of these and other aspects of physical oceanography see…

1. Density stratification in the ocean

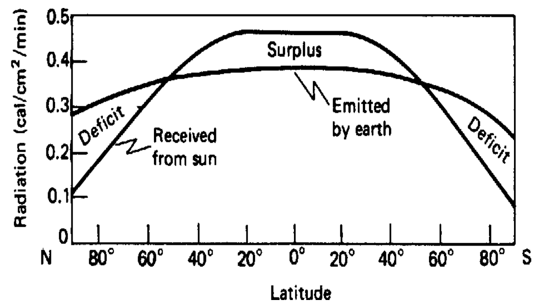
The density of seawater is controlled by its salt content or salinity and its temperature. Salinity is historically defined as the total salt content of seawater and the units were given as grams of salt per kilogram of seawater or parts per thousand (‰). Salinity was expressed on a mass of seawater basis because mass, rather than volume, is conserved as the temperature and pressure change. In modern oceanography salinity is determined as a conductivity ratio on the practical salinity scale and has no units (Millero, 1993). For more details on the formal definition of salinity and on the preparation of very accurate standards see the report by UNESCO (1981).

The salinity of surface seawater is controlled primarily by the balance between evaporation and precipitation. As a result the highest salinities are found in the so-called sub-tropical central gyre regions centered at about 20° to 30° North and South, where evaporation is extensive but rainfall is minimal. Surprisingly, they are not found at the equator where evaporation is also large, but so is rainfall. The highest surface salinities, other than evaporite basins, are found in the Red Sea.

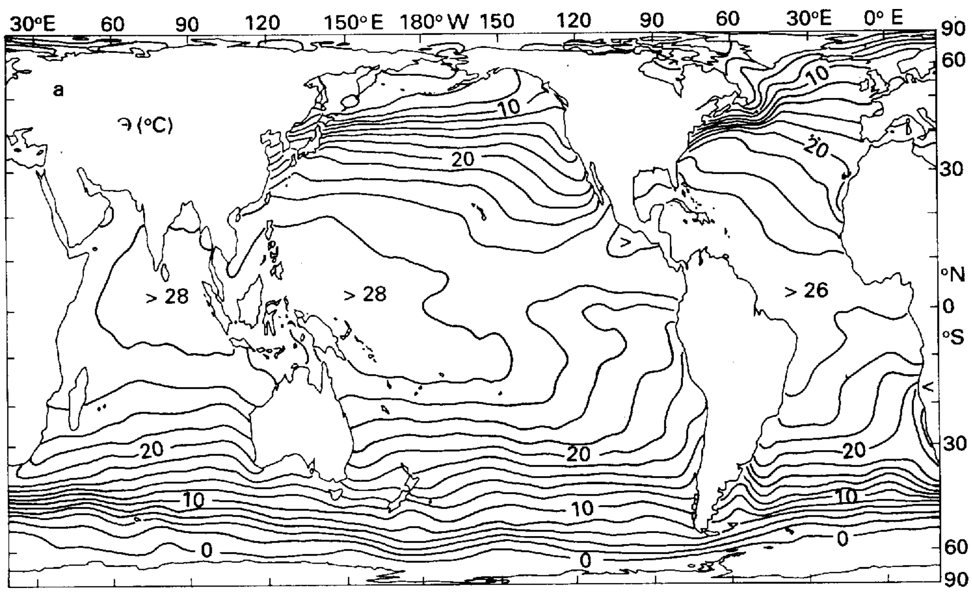




The temperature of seawater is fixed at the seasurface by heat exchange with the atmosphere. The average incoming energy from the sun at the earth's surface is about four times higher at the equator than at the poles. The average infrared radiation heat loss to space is more constant with latitude.

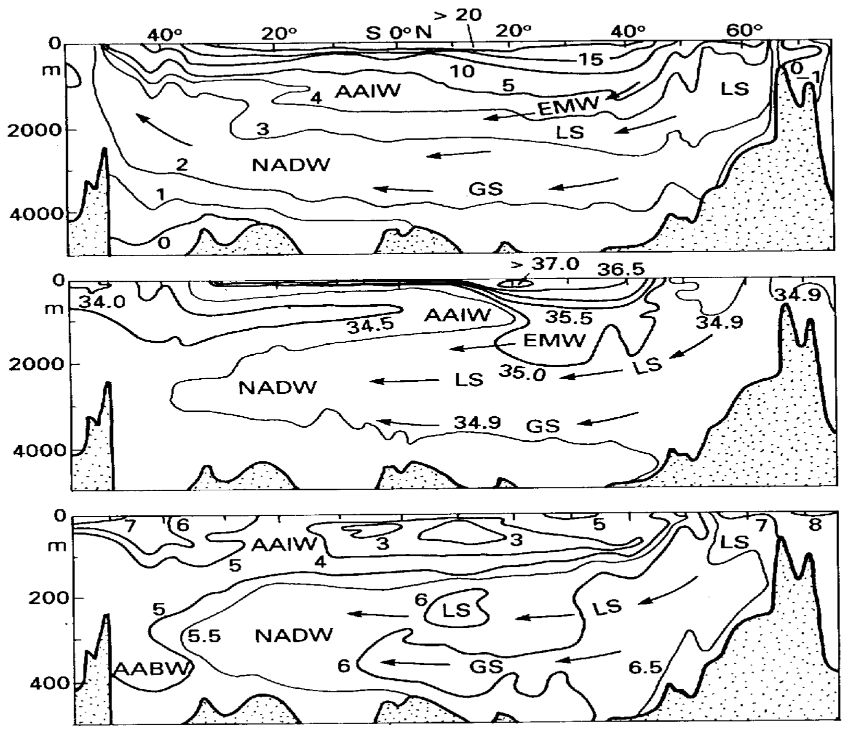


As a result there is a net input of heat to the earth's surface into the tropical regions and this is where we find the warmest surface seawater. Heat is then transferred from low to high latitudes by winds in the atmosphere and by currents in the ocean. The geothermal heat flux from the interior of the Earth is generally insignificant except in the vicinity of hydrothermal vents at spreading ridges (**references**) and in relatively stagnant locations like the abyssal northern North Pacific (Joyce, et al. 1986) and the Black Sea (Murray et al., 1991).

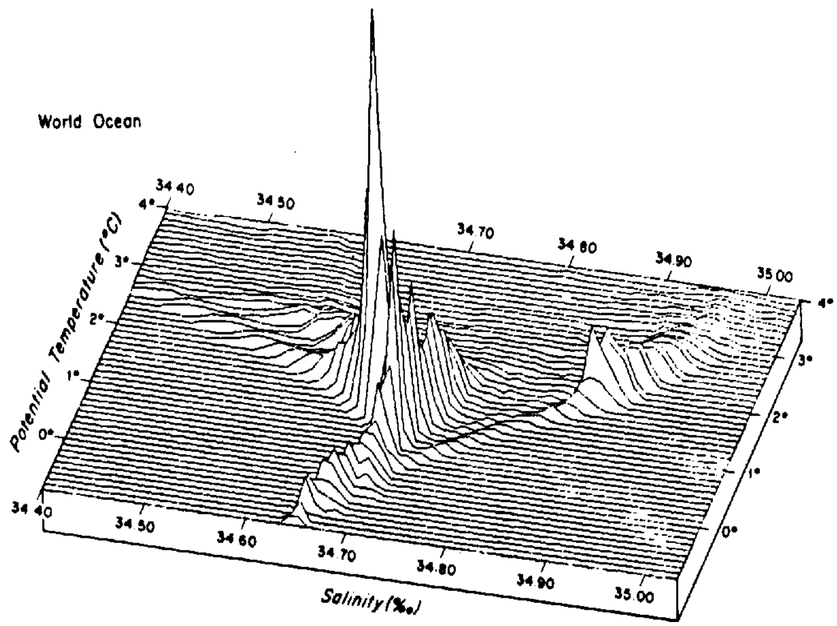


Because the seawater signatures of temperature and salinity are acquired by processes occurring at the air-sea interface we can also state that the density characteristics of a parcel of seawater are determined when it is at the seasurface. This density signature is locked into the water when it sinks. The density will be modified by mixing with other parcels of water but if the density signatures of all the end member water masses are known, this mixing can be unraveled and the proportions of the different source waters to a given parcel can be determined.

To a first approximation the vertical density distribution of the ocean can be described as a three layered structure. The surface layer is the region from the seasurface to the depth having a temperature of about 10°C. The transition region where the temperature decreases from 10°C to 4°C is called the thermocline. The deep sea is the region below the thermocline.



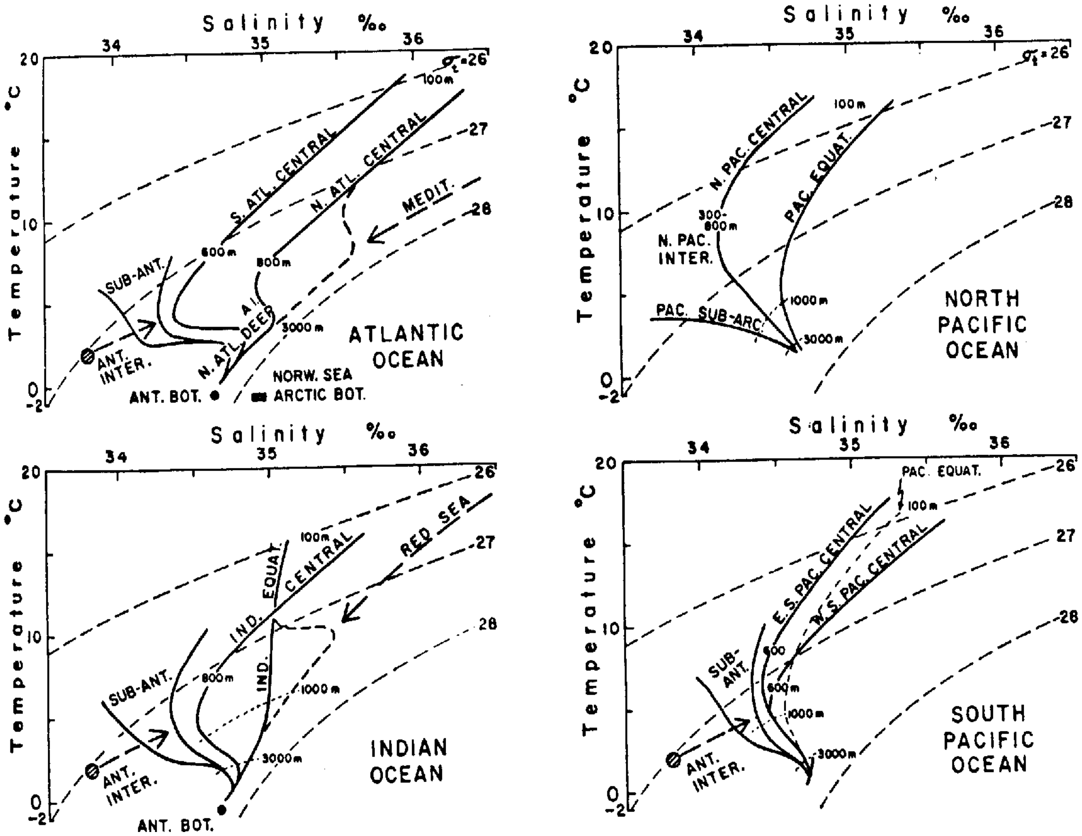
Because temperature (T) and salinity (S) are the main factors controlling density, oceanographers use T-S diagrams to describe the features of the different water masses. The average temperature and salinity of the world ocean and various parts of the ocean are given in Fig. B-4 and Table B-3.



The North Atlantic contains the warmest and saltiest water of the major oceans. The Southern Ocean (the region around Antarctica) is the coldest and the North Pacific has the lowest average salinity.

	Temperature (°C)	Salinity (‰)
Pacific (total)	3.14	34.60
North Pacific	3.13	34.57
South Pacific	3.50	34.63
Indian (total)	3.88	34.78
Atlantic (total)	3.99	34.92
North Atlantic	5.08	35.09
South Atlantic	3.81	34.84
Southern Ocean[b]	0.71	34.65
World ocean (total)	3.51	34.72

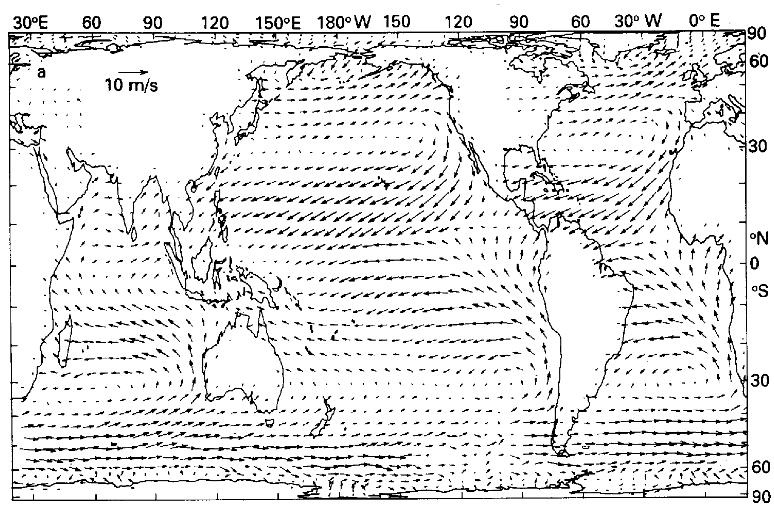
Conventional T-S diagrams for specific locations in the individual oceans are shown in Fig. B-5. The inflections in the curves reflect the inputs of water from different sources. The linear regions represent mixing intervals between these core sources. For example, in the Atlantic Ocean the curves reflect input from Antarctic Bottom Water(AABW), North Atlantic Deep Water (NADW), Antarctic Intermediate Water (AIW), Mediterranean Water (MW), and Warm Surface Water (WSW).

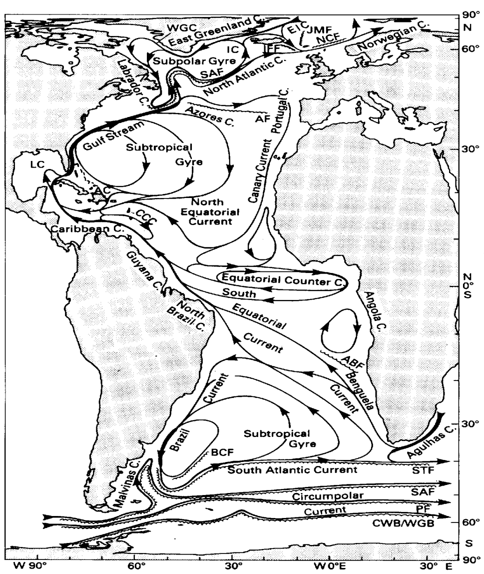


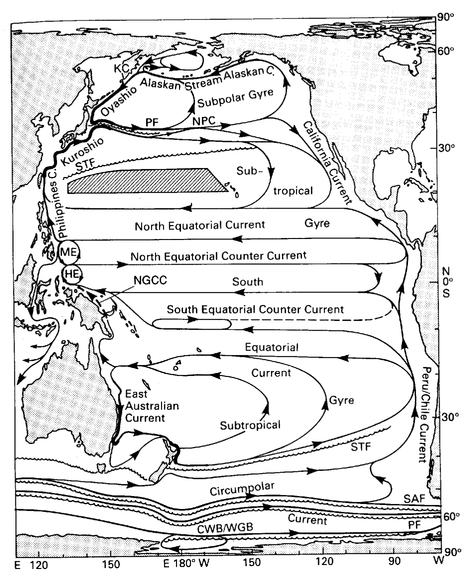
	This census			Montgomery (1958), Cochrane (1958), and Pollak (1958), modified to exclude adjacent seas		
	Volume (km³ × 10⁶)	Mean (°C)	Mean S (‰)	Volume (km³ × 10⁶)	Mean θ (°C)	Mean S (‰)
North Pacific	332.2	3.13	34.57	346.1	3.39	34.59
South Pacific	379.4	3.14	34.63	376.5	3.36	34.64
Pacific	711.6	3.14	34.60	722.6	3.37	34.62
Indian	282.7	3.88	34.78	291.6	3.70	34.75
Atlantic	326.2	3.99	34.92	321.8	3.76	34.87
World Ocean	1320.5	3.51	34.72	1336.0	3.54	34.71

2. Surface Currents

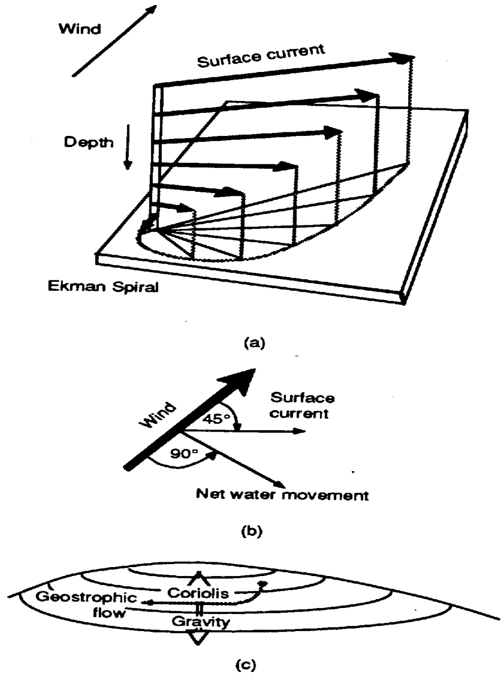
Surface ocean currents respond primarily to the climatic wind field. The prevailing winds supply much of the energy that drives surface water movements. This becomes clear when charts of the surface winds and ocean surface currents are superimposed. The *wind-driven circulation* occurs principally in the upper few hundred meters and is therefore primarily a horizontal circulation although vertical motions can be induced when the geometry of surface circulation results in convergences (downwelling) or divergences (upwelling). The depth to which the surface circulation penetrates is dependent on the water column stratification. In the equatorial region the currents extend to 30-500 m while in the circumpolar region where stratification is weak the surface circulation can extend to the sea floor.







The net direction of motion of the water is not always the same as the wind, because other factors come into play. These are shown schematically in Fig. B-6. The wind blowing across the seasurface drags the surface along and sets this thin layer in motion. The surface drags the next layer and the process continues downward, involving successively deeper layers. As a result of friction between the layers each deeper layer moves more slowly than the one above and its motion is deflected to the right (clockwise) in the northern hemisphere by the Coriolis force. If this effect is represented by arrows (vectors) whose direction indicates current direction and length indicates speed, the change in current direction and speed with depth forms a spiral. This feature is called the Ekman spiral. If the wind blew continuously in one direction for a few days a well developed Ekman spiral would develop. Under these conditions the integrated net transport over the entire depth of the Ekman spiral would be at 90° to the right of the wind direction (right in the northern hemisphere and left in the southern hemisphere). Normally the wind direction is variable so that the actual net transport is some angle less than 90°.



Ekman transport, changes in seasurface topography and the Coriolis force combine to form geostrophic currents. Take the North Pacific for example. The westerlies at ~40°N and the Northeast trades (~10°N) set the North Pacific Current and North Equatorial Current in motion as a circular gyre. Because of the Ekman drift, surface water is pushed toward the center of the gyre (~25°N) and piles up to form a seasurface "topographic high". As a result of the elevated seasurface, water tends to flow "downhill" in response to gravity. As it flows, however, the Coriolis force deflects the water to the right (in the northern hemisphere). When the current is constant and results from balance between the pressure gradient force due to the elevated seasurface and the Coriolis force, the flow is said to be in geostrophic balance. The actual flow is then nearly parallel to the contours of the elevated seasurface and clockwise. The seasurface topography of the Pacific Ocean was determined by Tai and Wunsch (1983) from satellite altimetry. The absolute elevation of the subtropical gyre can be clearly seen and fits the schematic description given above.

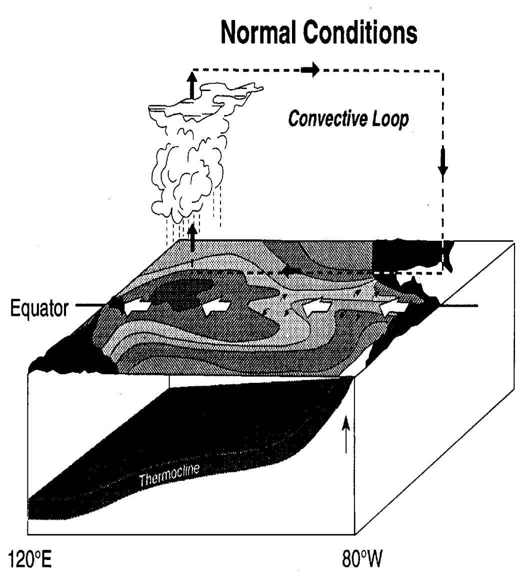
As a result of these factors (wind, Ekman transport, Coriolis force) the surface ocean circulation in the mid latitudes is characterized by clockwise gyres in the northern hemisphere and the counterclockwise gyres in the southern hemisphere. The main surface currents around these gyres for the world's oceans are shown in Fig. B-7. The regions where Ekman transport tends to push water together (such as the subtropical gyres) are called convergences. Divergences (such as the equator) result when surface waters are pushed apart.

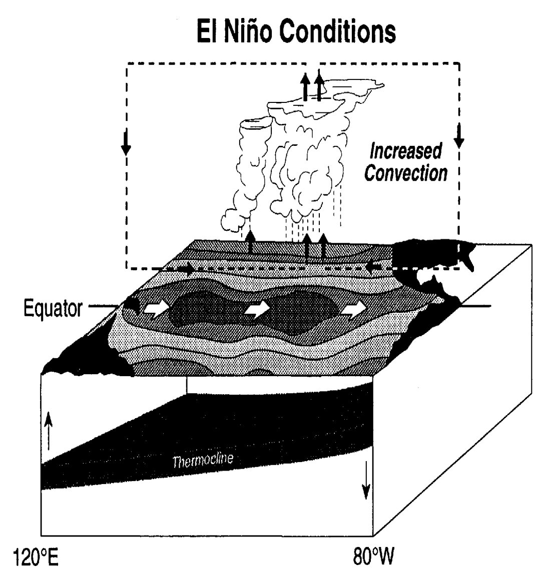
Total transport by the surface currents varies greatly and reflects the mean currents and cross sectional area. Some representative examples will illustrate the scale. The transport around the subtropical gyre in the North Pacific is about 70 Sv (1 Sverdrup or Sv = 1 x 10^6 m^3 s^{-1}). The Gulf Stream, which is a major northward flow off the east coast of North America, increases from 30 Sv in the Florida Straits to 150 Sv at 64°30'W, or 2000 km downstream.

3. El Nino-Southern Oscillation

The equatorial Pacific is one of the best studied regions of ocean divergence (Philander, 1990; McPhaden et al., 1998). This is because of the El Nino-Southern Oscillation phenomenon. The region around the equator normally experiences strong easterly trade winds that result in divergence from the equator and upwelling of colder, nutrient rich water from below (Fig. B-8a). This "cold-tongue" typically extends from the coast of South America to about the date line (180°). The trade winds also drive near-equatorial surface flow westward as the South Equatorial Current (SEC). This piles up warm surface water in the western Pacific to create a deep warm pool. As a result the thermocline has a tilt and is deeper in the west than the east. The westward flow in the surface SEC is partly compensated by a return flow to the east in the thermocline (~150m) called the Equatorial Undercurrent (EUC).

There is a zonal atmospheric circulation system associated with this normal ocean condition called the Walker Cell (Fig. B-8a). Evaporation rates are high over the warm pool and warm moist air ascends to great heights (deep convection) producing extensive cloud systems and rain. The Walker Cell is closed by westerly winds aloft and subsidence in the high-pressure zone of the eastern Pacific.

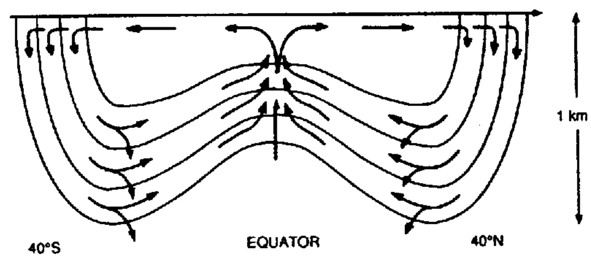




During El Nino (Fig B-8b) the trade winds weaken, and even reverse, in the central and western Pacific resulting in a local eastward acceleration of the surface currents. Westerly wind events in the western Pacific excite downwelling equatorial Kelvin waves which propagate into the eastern equatorial Pacific where they depress the thermocline (Kessler and McPhaden, 1995)(Fig B-8b). The winds in the eastern Pacific are usually still easterly favoring upwelling, but because the thermocline is depressed, warmer water is upwelled. The net result is migration of the "warm pool" and its associated atmospheric deep convection from the western Pacific to east of the date line (Fig B-8b). Anomalously warm sea surface temperatures occur from the coast of South America to the date line.

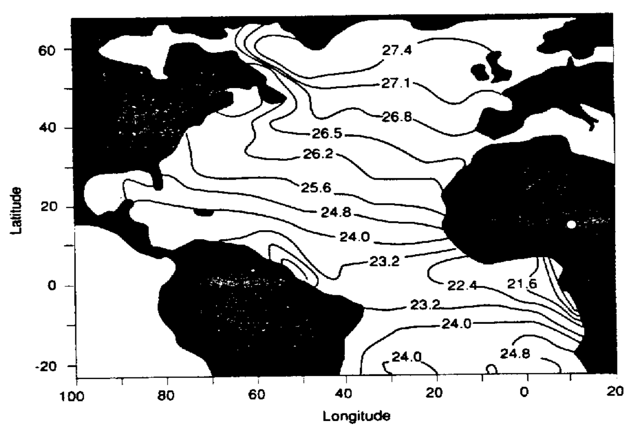
Deep Convergence in the atmosphere is the main driving force for atmospheric circulation through the release of latent heat at mid-tropospheric levels. The zonal shift in the site of deep convection during El Nino affects atmospheric circulation and climate on a global basis (Wallace et al., 1998). The variations in the upwelling also influence the flux of CO_2 from the ocean to the atmosphere (Feely et al., 1997) and the biological characteristics of the region (Murray et al., 1994).

4. Thermocline Circulation



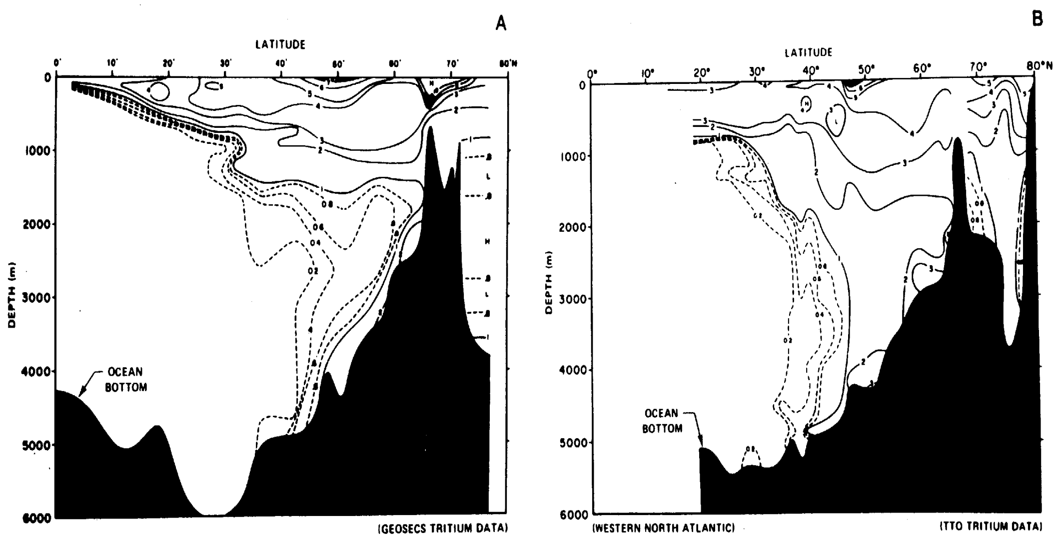
The transition region between the surface and deep ocean is referred to as the thermocline. This is also a pycnocline zone where the density increases appreciably with increasing depth. Most of the density change results from the decrease in temperature (hence thermocline).

A simple but physically realistic model based on lateral transport has evolved to explain the origin of the thermocline. According to this view, the interior of the ocean is ventilated by rapid mixing and advection <u>along</u> isopycnal surfaces (Jenkins, 1980). The density surfaces that lie in the thermocline at 200 to 1000 m in the equatorial and mid-latitude regions shoal and outcrop at high latitudes. The argument is that water acquires its T and S (and chemical tracer) signature while at the sea surface and then sinks and is transported horizontally as shown in Fig. B-9. A map showing the winter outcrops of isopycnal surfaces in the North Atlantic Ocean is shown in Fig. B-10. Characteristic values of the horizontal eddy diffusion coefficient (K) are of the order of 10^7 cm^2/s. Assuming a distance (L) of the order of 2000 km (30°) and assuming the characteristic time is $\tau = L^2/K$, we obtain a characteristic ventilation time for the main thermocline of about 130 years.



The horizontal isopycnal thermocline model is important for the problem of determining the fate of the excess atmospheric CO_2. The increase of CO_2 in the atmosphere is modulated by transport of excess CO_2 from the atmosphere into the interior of the ocean. The direct ventilation of the thermocline in its outcropping regions at high latitudes plays an important role in removing CO_2 from the atmosphere (Siegenthaler and Sarmiento, 1993).

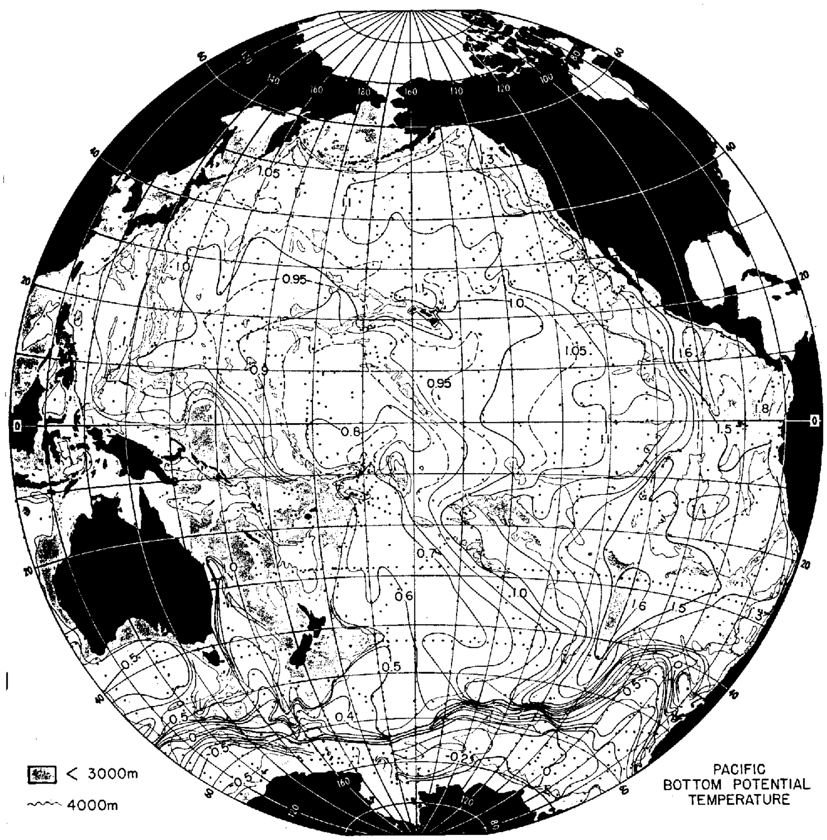
Nuclear bomb produced $^{14}CO_2$ and ^{3}H (as HTO) have been used to describe and model this rapid thermocline ventilation (Ostlund *et al.*, 1974; Sarmiento *et al.*, 1982; Fine *et al.*, 1983). For example, changes in the distributions of tritium in the western Atlantic between 1972 (GEOSECS) and 1981 (TTO) are shown in Fig. B-11 (Baes and Mulholland, 1985). In the 10



years following the atmospheric bomb tests of the early 1960's, a massive penetration of ^{3}H into the thermocline at all depths has occurred. Comparison of the GEOSECS and TTO data, which have a 9 year time difference, clearly shows the rapid ventilation of the North Atlantic and the value of such "transient" tracers. A similar distribution can be seen in the distribution of man-made chlorofluorocarbons, which have been released over a longer period (40 years) (Gammon *et al.*, 1982).

5. Abyssal Circulation

The circulation of the deep ocean below the thermocline is referred to as abyssal circulation. The currents are slow and difficult to measure but the pattern of circulation can be clearly seen in the properties of the abyssal water. For example, the water of lowest temperature in the water column is usually the densest and lies deepest. As a result, charts of the bottom water temperature have been useful in describing the pattern of the abyssal circulation (*e.g.*, Mantyla, 1975; Mantyla and Reid, 1983). The topography of the sea floor plays an important role in constraining the circulation and much of the abyssal flow is funneled through passages such as the Denmark Straight, Gibbs Fracture Zone, Vema Channel, Samoan Passage, and Drake Passage.



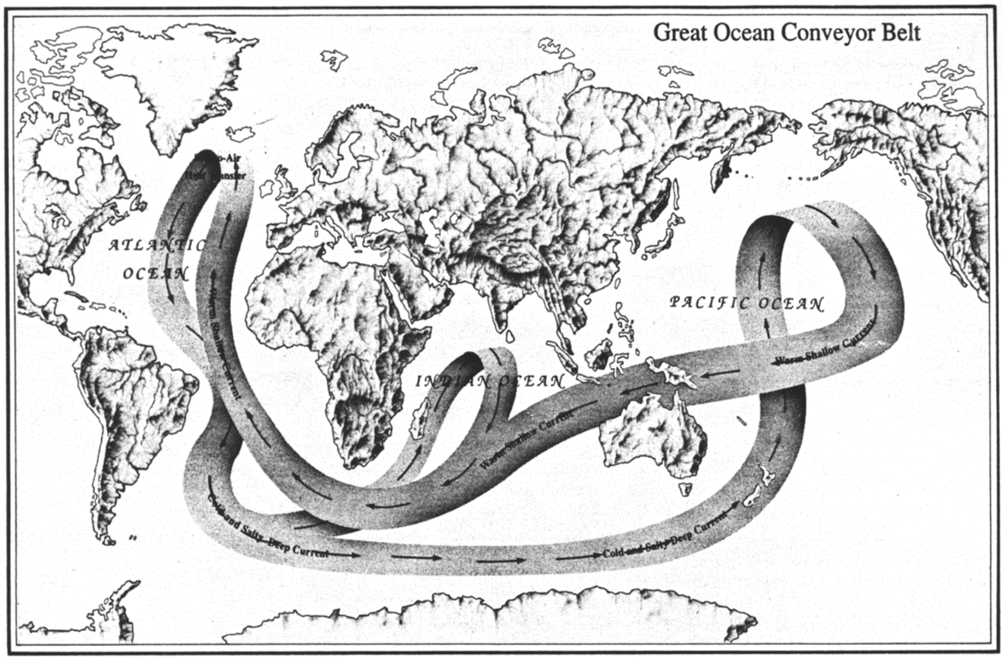
For a steady state ocean, a requirement of the *heat balance* is that the input of new cold abyssal water (Antarctic Bottom Water and North Atlantic Deep Water) sinking in the high latitude regions must be balanced by the input of heat by geothermal heating (heat flow from the Earth), downward convection of relatively warm water (*e.g.,* from the Mediterranean) and downward diffusion of heat across the thermocline. A general mass balance of the world's oceans requires that the water sinking in the polar regions must be exactly balanced by the upwelling of water from the abyssal ocean to the surface water. A combination of the mass and heat balances together with the forcing of the wind and the effect of a rotating Earth determine the nature of the abyssal circulation.

6. The Ocean Conveyor Belt

The ocean conveyor-belt is one of the major elements of today's ocean circulation system (Broecker, 1997). A key feature is that it delivers an enormous amount of heat to the North Atlantic and this has profound implications for past, present and probably future climates.

The conveyor-belt is shown schematically in Fig. B-12. Warm and salty surface currents in the western North Atlantic (e.g. the Gulf Stream) transport heat to the Norwegian-Greenland Seas where it is transferred to the atmosphere. This heat helps moderate the climate of northern Europe. The cooling increases the density resulting in formation of the now cold and salty North Atlantic Deep Water (NADW)(Worthington, 1970). The NADW travels south through the North and South Atlantic and then joins the Circumpolar Current that travels virtually unimpeded in a clockwise direction around the Antarctic Continent.

Deep water also forms along the margins of Antartica and feeds the Circumpolar Current. The Weddell Sea, because of its very low temperature, is the main source of Antarctic Bottom Water (AABW) which flows northward at the very bottom into the South Atlantic, and then through the Vema Channel in the Rio Grande Rise into the North Atlantic. It ultimately returns southward as part of the NADW.



The circumpolar current is a blend of waters of North Atlantic (~47%) and Antarctic margin (~53%) origin (Broecker, 1997). This current is referred to as the Pacific Common Water and is the source of deep water to the Indian and Pacific Oceans. Deep water does not form in a similar way in the North Pacific because the salinity is too low (Warren, 1983). Pacific Common water enters the Pacific in the southwest corner and flows north along the western boundary of the Tonga Trench. The abyssal circulation model of Stommel (1958) and Stommel and Arons (1960) predicted that deep waters flow most intensely along the western boundaries in all oceans and gradually circulate into the interior with a cyclonic flow as allowed by topography. Most of the northward abyssal flow passes from the southwest Pacific to the north central pacific through the Samoan Passage, located west of Samoa. In the North Pacific, the abyssal flow splits and goes west and east of the Hawaiian Islands. These flows meet again north of Hawaii where they mix, upwell and flow back to the South Pacific at mid-depths.

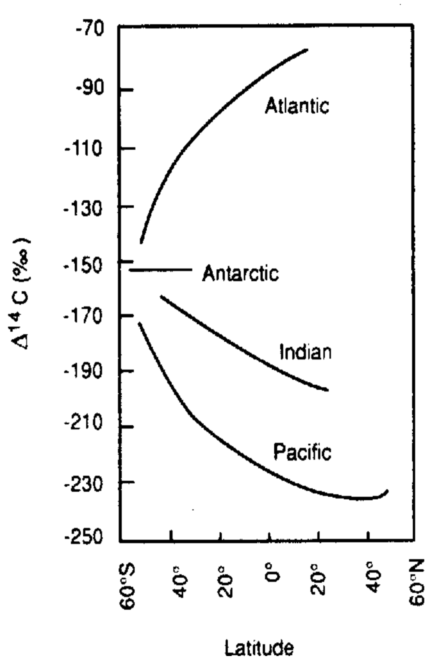
The conveyor-belt is completed by return flow of surface water from the Pacific to the Atlantic. There are two main paths of this return flow, which amounts to about 19

Sv. Some passes through the Indonesian Archepelgo, the Indian Ocean and around the tip of South Africa via the Agulhas Current (Gordon, 1985). Some enters the South Atlantic via the Drake Passage. Based on the salt budget, Broecker (1991) argued that the Drake Passage route transports about 50% more than the Agulhas Current. Finally there is a small transport (about 1 Sv) from the Pacific to the Atlantic through the Bering Strait that is important for the Arctic salt balance.

The salt budget for the Atlantic, which is determined in part by the flux of fresh water through the atmosphere, drives the conveyor belt and can explain how it has varied in the past. At present there appears to be a net water vapor loss of about 0.32 Sv (greater than the flow of the Amazon) from the Atlantic to the Pacific. The NADW transports about 16.3 Sv of water with a salinity of 34.91. This is produced from 15 Sv of Gulf Stream water with a salinity of 35.8, 1 Sv of transport from the Bering Straits with S = 32 and a net excess of river inflow and rainfall over evaporation of about 0.3 Sv (Zauker and Broecker, 1992). It is easy to show that small changes in the freshwater budget can have a significant impact. For example, if the excess of precipitation plus runoff over evaporation increased by 50% to 0.45 Sv, the salt content of the NADW would decrease to 34.59. In order to compensate for the resulting reduction in density the water would have to be cooled by an additional 1.4°C and the conveyor would have to more than double its flow to restore the salt balance (Broecker, 1997). Model simulations have also shown that the oceans' thermohaline circulation is extremely sensitive to freshwater input (Manabe and Stauffer 1995).

Although the general abyssal circulation patterns are fairly well known, it is difficult to quantify the rates of the various flows. Abyssal circulation is generally quite slow and variable on short time scales. The calculation of the rate of formation of abyssal water is also fraught with uncertainty. Probably the most promising means of assigning the time dimension to oceanic processes is through the study of the distribution of radioactive chemical tracers. Difficulties associated with the interpretation of radioactive tracer distributions lie both in the models used, non-conservative interactions, and the difference between the time scale of the physical transport phenomenon and the mean life of the tracer.

An example of the power of such tracers is in the "dating" of abyssal water using ^{14}C. ^{14}C has an atmospheric source and a half-life of 5720 years. Stuiver *et al.* (1983) measured the ^{14}C distribution in dissolved inorganic carbon in deep samples from major ocean basins (Fig. B-13). This data was used to calibrate a box model which indicated that the replacement times for Atlantic, Indian and Pacific Ocean deep waters (depths > 1500 m) are 275, 250, and 510 years respectively.



The present form of the conveyor belt appears to have been initiated by closure of the Panamanian seaway between the North and South American continents (Keigwin, 1982; Maier-Reimer et al., 1990). Geological evidence indicates that gradual closing of the Isthmus of Panama lasted from 13 to 1.9 Myr ago. Paleoceanographic records indicate that closure was sufficient by 4.6 Myr ago to cause a marked reorganization of ocean circulation (Burton et al., 1997; Haug and Tiedemann, 1998). At this time the Gulf Stream intensified resulting in the transport of warm water to high latitudes. As a result NADW formation intensified and increased atmospheric moisture input to high latitudes which probably helped trigger the growth of northern hemispheric ice-sheets.

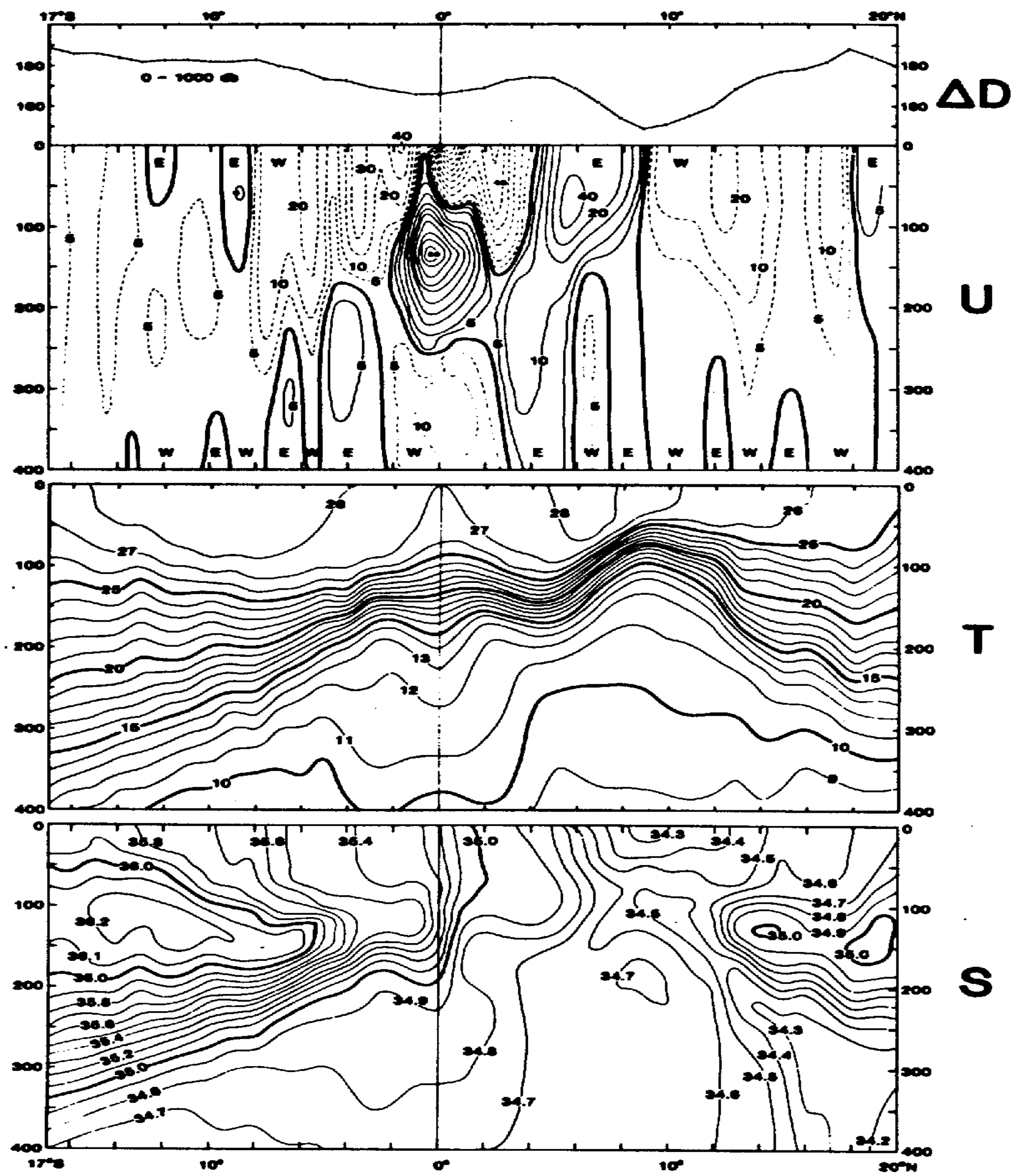
There is strong evidence that the conveyor belt has switched regularly from one mode of operation to another in the past. The associated changes in climate have been large, abrupt and global (Denton and Hendy, 1994). The changes appear to be driven by factors controlling the density of high-latitude North Atlantic surface water. These events appear to have been triggered by an increase in ice berg input, mainly from Canada (Bond et al., 1992). These ice bergs transport terrigenous debris across the North Atlantic. When they melt they deposit a layer if ice rafted material on the sea floor. These periodic events in the geological record are called Heinrich events (Broecker, 1994; Bond et al., 1997). The input of fresh water reduces production of NADW thus shutting down the present mode of the conveyor belt. The timing of these events has been perfectly preserved in the sediments from as far away as the Santa Barbara Basin (Behl and Kennett, 1996) and the glaciers in the Columbian Andes (Thompson et al, 199x). At the time of these event the climate cools at high latitudes and globally.

The climate records in Greenland ice reveal that over the past 60 kyr conditions switched back and forth between intense cold and moderate cold on a few 1000 yr time scale. These so-called Dansgaard-Oeschger cycles are characterized by abrupt changes in temperature, dust content, ice accumulation rate, methane concentration and CO_2 content. The onset of these cold events occurred on time scales as short as a few decades to a few years (Alley et al., 1993). Each period of intense cold has been matched by an ice-rafting or Heinrich event in North Atlantic sediments. As a result of the switch to a colder climate, ice-berg production slows and the salinity of the north Atlantic surface water slowly increases enabling NADW formation to occur again. The return to the warm phase occurs much more slowly, over a 1000 yr time frame. These cyclic events appear to have continued in the Holocene, although with much muted amplitude (Alley et al., 1997).

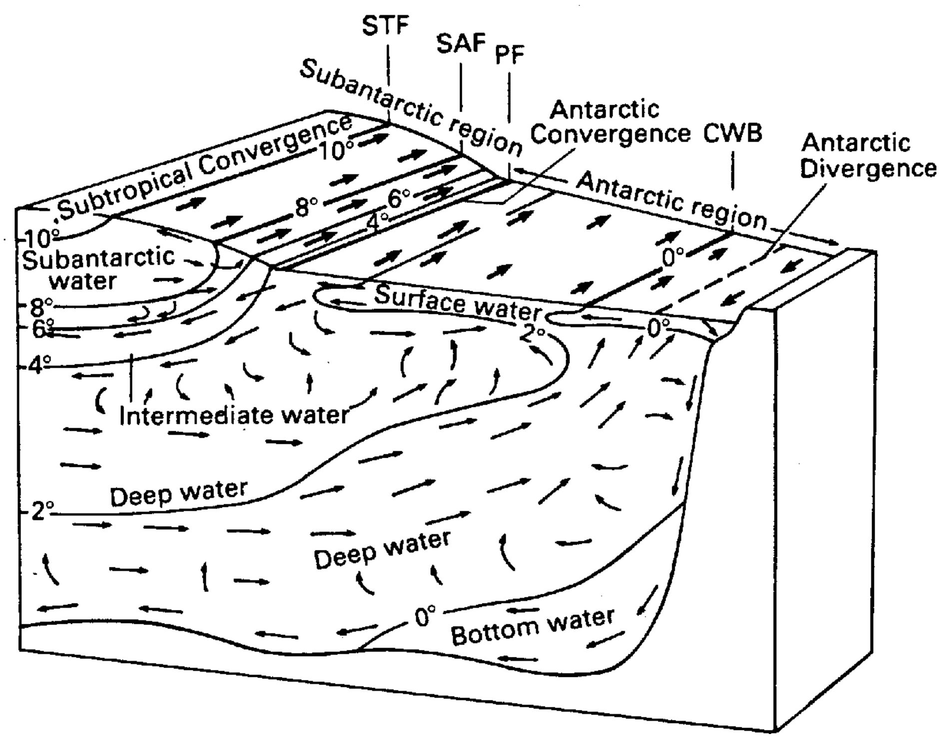
There is great concern that one of the effects of global warming could be reduction in formation of NADW and associated reorganization of the conveyor-belt circulation (Manabe and Stouffer, 1995). The consequences of global warming will be to warm surface seawater and to intensify the hydrological cycle. Both factors will make it more difficult to form deep water and could lead to an “anthropogenically” induced climate shift. Paradoxically, global warming could result in climate cooling for northern Europe.

Detailed Structure: Equatorial Pacific

Descriptive Physical Oceanography



Detailed Structure: Antarctic Polar Front



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