**Lecture 3: Temperature, Salinity, Density and Ocean Circulation**
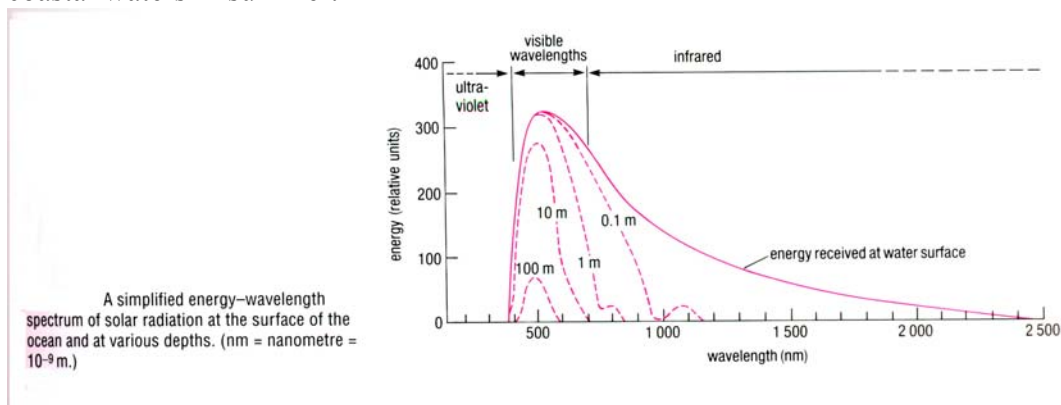
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Two of the most important characteristics of seawater are temperature and salinity – together they control its density, which is the major factor governing the vertical movement of ocean waters.
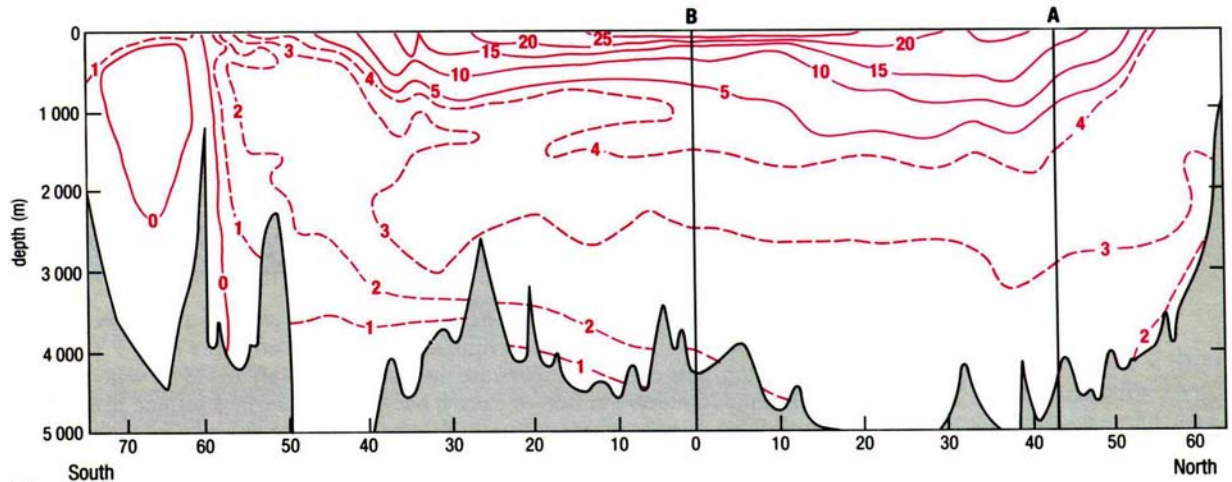
Temperature Distribution in the Ocean

The temperature of seawater is fixed at the sea surface 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 and in relatively stagnant locations like the abyssal northern North Pacific (Joyce, et al. 1986) and the Black Sea (Murray et al., 1991).
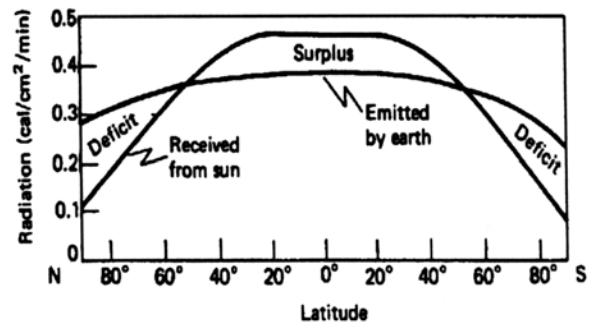
Water is transparent, so the radiation penetrates some distance below the surface; heat is also carried to deeper levels by mixing. Due to the high specific heat of water, diurnal and seasonal temperature variations are relatively small compared to the variations on land; oceanic temperature variations are on the order of a few degrees, except in very shallow water. Most solar energy is absorbed within a few meters of the ocean surface, directly heating the surface water and providing the energy for photosynthesis by marine plants and algae. Shorter wavelengths penetrate deeper than longer wavelengths. Infrared radiation is the first to be absorbed, followed by red, and so on. Heat conduction by itself is extremely slow, so only a small proportion of heat is transferred downwards by this process. The main mechanism to transfer heat deeper is turbulent mixing by winds and waves, which establishes a **mixed surface layer** that can be as thick as 200-300 meters or even more at mid-latitudes in the open ocean in winter or less than 10 meters in sheltered coastal waters in summer.

A simplified energy–wavelength spectrum of solar radiation at the surface of the ocean and at various depths. (nm = nanometre = 10^{-9} m.)

Between about 200 m and 1000 m depth, the temperature declines rapidly throughout much of the ocean. This region of steep temperature gradient is known as the **permanent thermocline**, beneath which, from about 1000 m to the ocean floor, there is virtually no seasonal variation and the temperatures are around 2 °C. This narrow range is maintained throughout the deep oceans, both geographically and seasonally, because it is determined by the temperature of the cold, dense water that sinks at the polar-regions and flows towards the Equator. Vertical distribution of temperature in the deep ocean is controlled by density driven water movements.

(a)

A vertical section showing the mean distribution of temperature in the western Atlantic Ocean, and two temperature depth profiles corresponding to locations A and B.

Contours of equal temperature are called **isotherms**
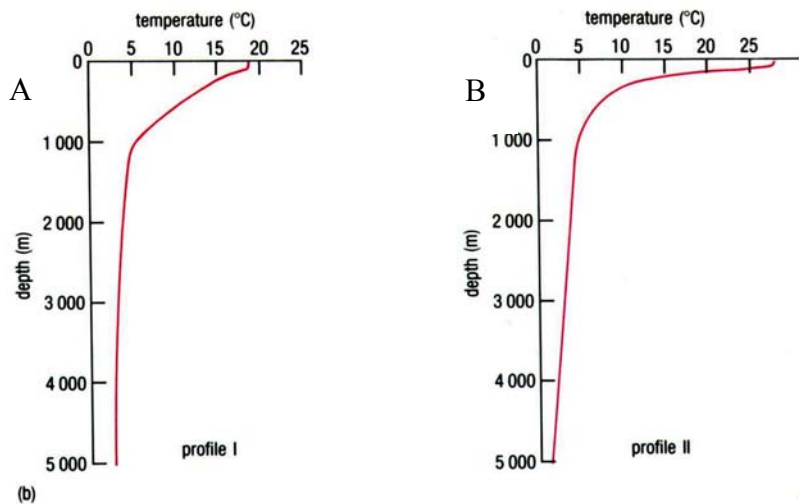
(b)

Above the permanent thermocline, the distribution of temperature with depth shows seasonal variations. In mid latitudes a **seasonal thermocline** often starts to develop in the spring above the permanent thermocline, as surface temperature rise and mixing by wind is small, in summer this seasonal thermocline reaches the maximum development (steepest gradient). In places or during periods of extensive mixing the mixed surface layer may extend to the permanent thermocline.

Bear in mind that while the locations of mean isotherms in the ocean do not change significantly on time scales of decades, the structure is maintained dynamically. In other words, the temperature at any particular location and depth – at least below the mixed layer – changes very little from year to year, even though the actual water at that location is changing all the time.

Horizontal advection in the ocean (surface currents) transports warm waters to colder regions and vice versa. Relatively warm water is carried to higher latitudes along the western sides of the ocean basins and cool water flows from high to low latitudes in the eastern margins.

Salinity Distribution in the Ocean
(Chapter 3 from Pilson (1998) has a good presentation)

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. The highest surface salinities, other than evaporite basins, are found in the Red Sea.

Salinity (S) conceptually ≡ grams of dissolved (<0.5 μm) inorganic ions per kg of seawater. The average salinity of seawater is S = 35 which means that SW is 3.5% salt and 96.5% H_2O by weight.

Why is salinity important?
1. Salinity, along with temperature, determines the density of seawater, and hence its vertical flow patterns in **thermohaline** circulation.
2. Salinity records the physical processes affecting a water mass when it was last at the surface.
    a. precipitation/evaporation – salts excluded from vapor
    b. freezing/thawing – salts excluded from ice
3. Salinity can be used as a **conservative** (unchanging) tracer for determining the origin and mixing of water types.

The **major ions** (>1mg/kg seawater) at S = 35.000 (from Pilson)

Ion	Formula	g/Kg	mmol/Kg
Sodium	Na^+	10.781	468.96
Magnesium	Mg^{2+}	1.284	52.83
Calcium	Ca^{2+}	0.4119	10.28
Potassium	K^+	0.399	10.21
Strontium	Sr^{2+}	0.00794	0.0906
Chloride	Cl^-	19.353	545.88
Sulfate	SO_4^{2-}	2.712	28.23
Bicarbonate	HCO_3^-	0.126	2.06
Bromide	Br^-	0.067	0.844
Borate	$H_3BO_4^-$	0.0257	0.416
Fluoride	F^-	0.00130	0.068
Totals	11	35.169	1119.87

Weight% of Major Ions
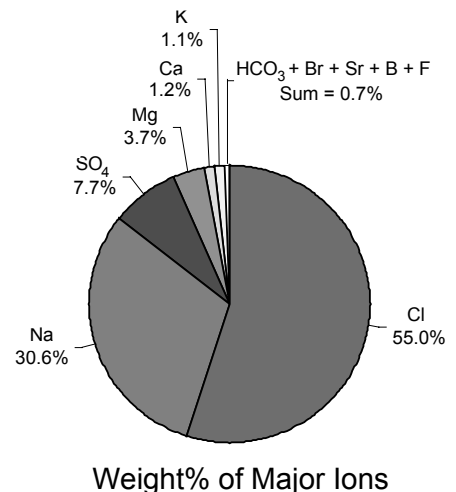
Historical perspective:

Marcet (1819) – determined that the major elements in seawater from six different areas are present in constant proportions to each other. This is now called the Marcet Principle.
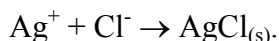
Dittmar (1884) – analyzed 77 samples collected from various depths throughout the world's oceans, during the cruise of the H.M.S. Challenger (1873-1876). This is considered as the first analyses of the major elements in seawater.

The "**Law of Constant Relative Proportions**" The ratios of the **major** ions in seawater are constant - with slight exceptions for HCO_3^- ($\pm<20\%$), Ca^{2+} ($\pm<1\%$) and Sr^{2+} ($\pm<2\%$).
  1. Thus, $[Na^+]/[Cl^-]$ is the same for all seawater (Atlantic or Pacific, surface or deep).
  2. The ratio of any major ion to salinity (e.g. $[Cl^-]/S$) is also constant, or nearly so. Ion concentrations in different waters are often normalized to salinity for comparisons.

How is Salinity Determined? (see Millero (1996) Chapter 2)

1. Boiling seawater, so that residual salts can be weighed, **does not work**. Volatiles are lost in the process (HCl, CO_2, Br and I); High temperatures are required to drive off water from hygroscopic salts (e.g. from $CaCl_2 \bullet xH_2O$). You always come up "light".
   a. bromide (and iodide) salts are volatile at high temperatures (~500-600 °C) and lost
   b. $MgCO_{3(s)}$ decomposes to MgO and CO_2 and $Ca(HCO_3)_2$ makes $CaO_{(s)} + H_2O + 2CO_2$. The CO_2 gas is lost
   c. $CaCl_2 \bullet xH_2O$ and $MgCl_2 \bullet xH_2O$ both decompose to give $HCl_{(g)}$ which is lost.

2. Analyze all major ions and sum. Problems: 11 separate analyses, and some are difficult to make, lots of work, cumulative errors are large.

3. Sorensen's Gravimetric Method – Add HCl and Cl_2, heat to 480 °C and weigh salt residue:
   a. reaction $= Cl_2 + 2Br^- \rightarrow 2Cl^- + Br_2$ (the latter is volatile, as is I_2 formed from I^-)
   b. This reproducible procedure corrects for loss of Br and I salts (converts them quantitatively to Cl_2 before they are lost), but there is no correction for CO_2 loss.

3. Sorensen's Titration Method – Used because the gravimetric method is a pain. Utilize the "law of constant proportions" or "Marcet's Principle"; Chlorinity approach (Cl%); Measure Cl % (which actually gives the sum of all halides; $Cl^- + F^- + Br^- + I^-$) by titration with $AgNO_3$ to precipitate AgCl, AgBr etc.. The titration gives the grams of Cl equivalent in 1 kg seawater.
$$Ag^+ + Cl^- \rightarrow AgCl_{(s)}.$$
   Chlorinity (Cl) $\equiv$ mass of chlorine equivalent to total mass of halogen in 1 Kg seawater. Salinity $= 1.80655$ Cl. This approach is fast, precise and not a bad approach.

4. The modern approach is to measure salinity in a **salinometer** by **conductance**.
   a. sample conductivity is determined as a ratio to a standard of KCl in pure water (at 15 °C and 1 atm pressure). Accordingly, salinity has no units using this method.
   b. **"practical salinity"** is calculated for a seawater sample using a polynomial expression in terms of the conductivity ratio (R), the actual temperature (t) of the sample, and a flock of fitting constants (a, b and k):

$$S = a_o + a_1R^{1/2} + a_2R + a_3R^{3/2} + a_4R^2 + a_5R^{5/2} + \frac{t-15}{1+k(t-15)} (b_o + b_1R^{1/2} + b_2R + b_3R^{3/2} + b_4R^2 + b_5R^{5/2})$$

   c. The typical precision of this method is 1/40,000, equivalent to ~0.001.

Surface seawater salinities largely reflect the local balance between evaporation and precipitation.
a. Low salinities occur near the equator due to rain from rising atmospheric circulation.
b. High salinities are typical of the hot dry gyres flanking the equator (±20-30 degrees latitude) where atmospheric circulation cells descend.
c. Salinity can also be affected by sea ice formation/melting (e.g. around Antarctica)
d. The surface N. Atlantic is saltier than the surface N. Pacific, making surface water denser in the N. Atlantic at the same temperature and leading to **down-welling** of water in this region this difference is because on average N. Atlantic is warmer (10.0 °C) than N. Pacific (6.7 °C). This is mostly because of the greater local heating effect of the Gulf Stream, as compared to the Kuroshio Current. Warmer water evaporates more rapidly, creating a higher residual salt content
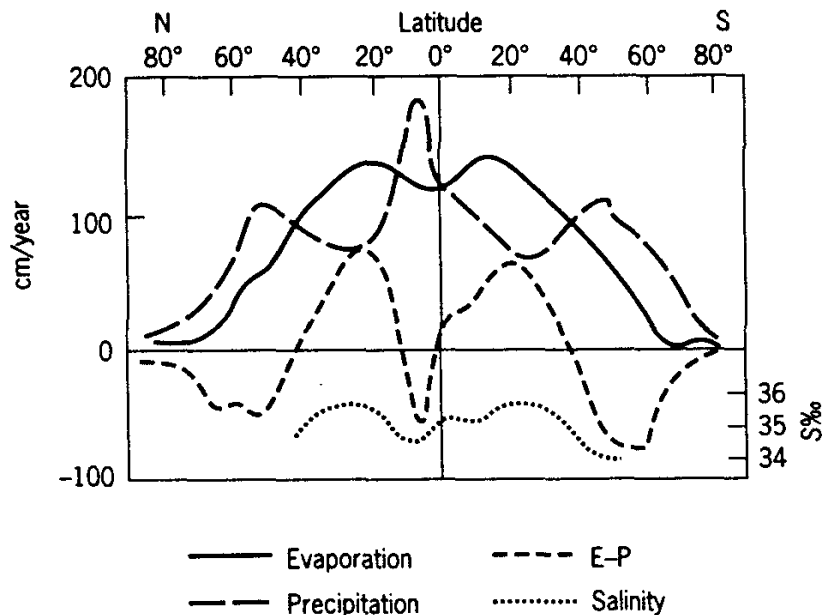
The influence of surface fluctuations in salinity due to changes in evaporation and precipitation is generally small below 1000 m, where salinities are mostly between about 34.5 and 35.0 at all latitudes. Zones where salinity decreases with depth are typically found occur at low latitudes and mid latitudes, between the mixed surface layer and the deep ocean. These zones are known as **haloclines**.
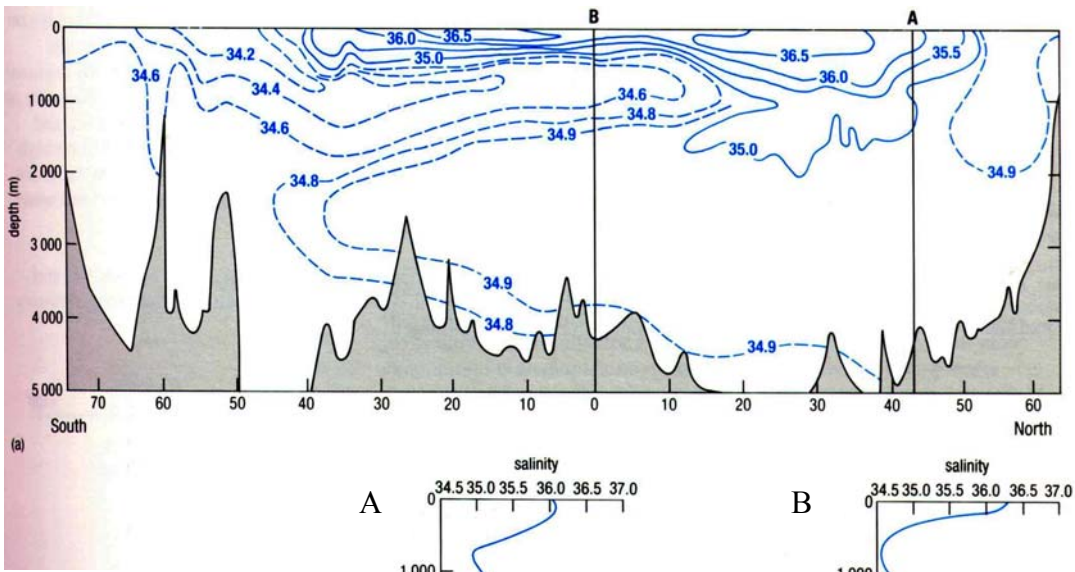
A vertical section showing the mean distribution of salinity in the western Atlantic Ocean, and two salinity depth profiles corresponding to locations A and B.
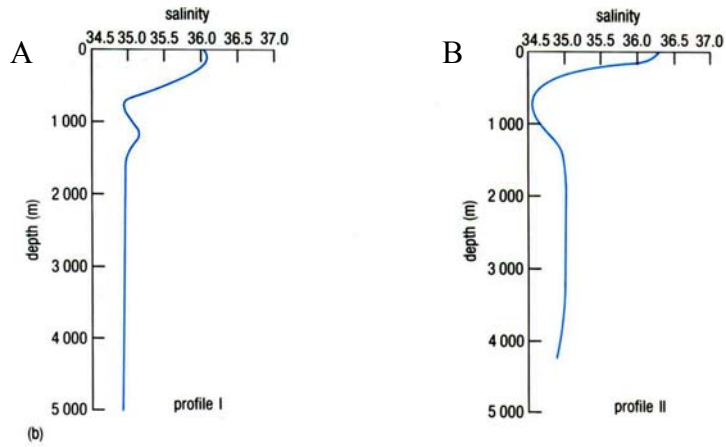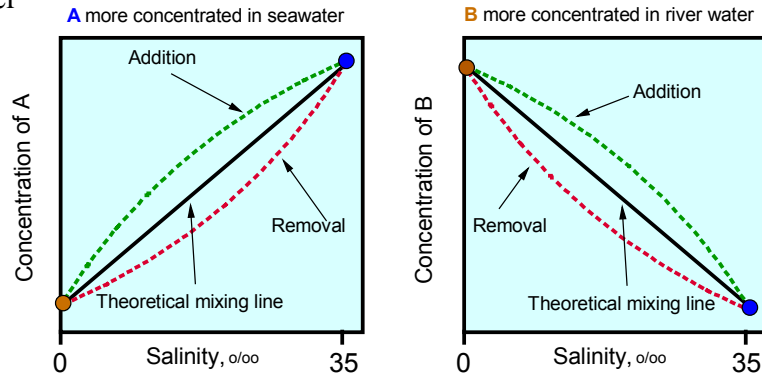
Salinity distribution in surface waters (isohaline surfaces)
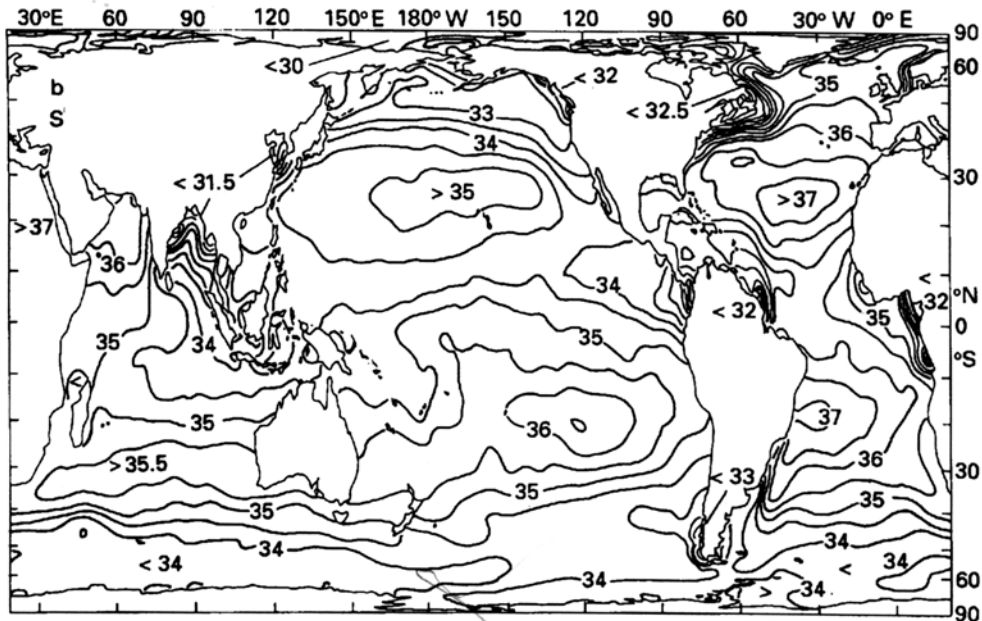
Salinity as a Conservative Tracer for Water Mixing in Estuaries

As seawater mixes with river water in estuaries, a conservative seawater component (e.g. a major ion) will co-vary directly in concentration with salinity. A component that is added during mixing will curve up in concentration versus a conservative mixing line (yielding an up "banana plot")

A more concentrated in seawater

Addition

Removal

Theoretical mixing line

Concentration of A

Salinity, o/oo   0 — 35

B more concentrated in river water

Addition

Removal

Theoretical mixing line

Concentration of B

Salinity, o/oo   0 — 35

A component that is lost during mixing will curve down in concentration versus a conservative mixing line (yielding a down "banana plot") as shown in the figures above. Such non-conservative behavior will only be evident if the addition or loss process is relatively rapid versus water mixing in the estuary.

Seawater Density

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 sea surface. Temperatures of seawater vary widely (-1 to 30 °C), whereas the salinity range is small (35.0 ± 2.0). 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.

Water Mass	Temp (°C)	Salinity
North Atlantic Central Water	8-19	35.1-36.5
Antarctic Circumpolar Water	0-2	34.6-34.7
Antarctic Intermediate Water	3-7	33.8-34.7
North Pacific Intermediate Water	4-10	34.0-34.5
North Atlantic Deep Water	2-4	34.8-35.1
Antarctic Bottom Water	-0.4	34.7

This density signature is locked into the water parcel 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.
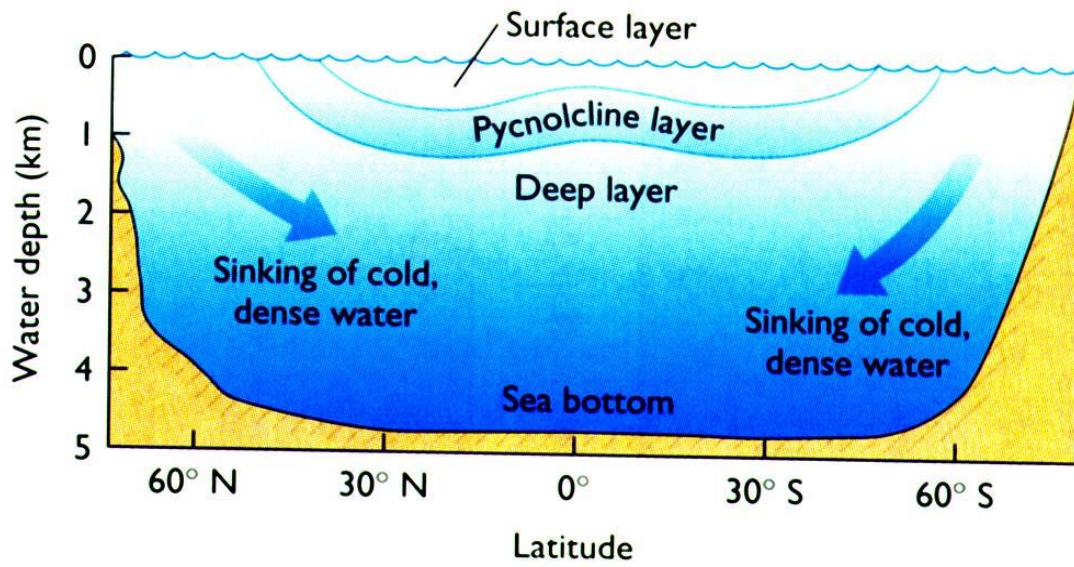
DENSITY STRUCTURE OF THE OCEANS
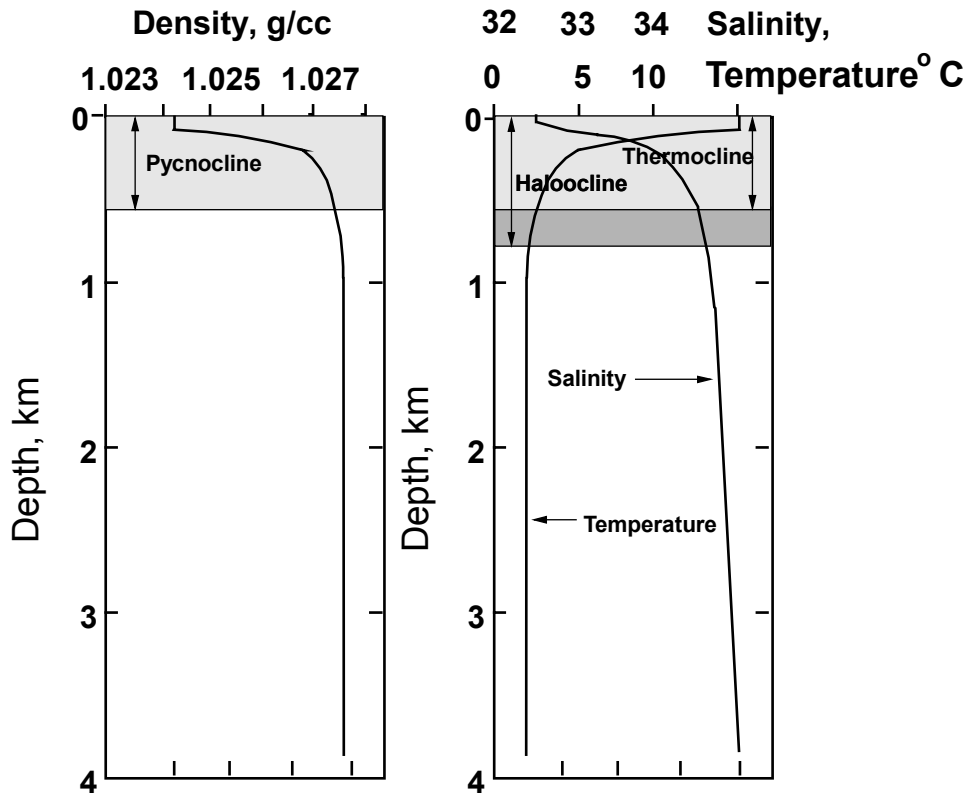
Because temperature (T) and salinity (S) are the main factors controlling density, oceanographers use T-S diagrams to describe the features of the different water masses.
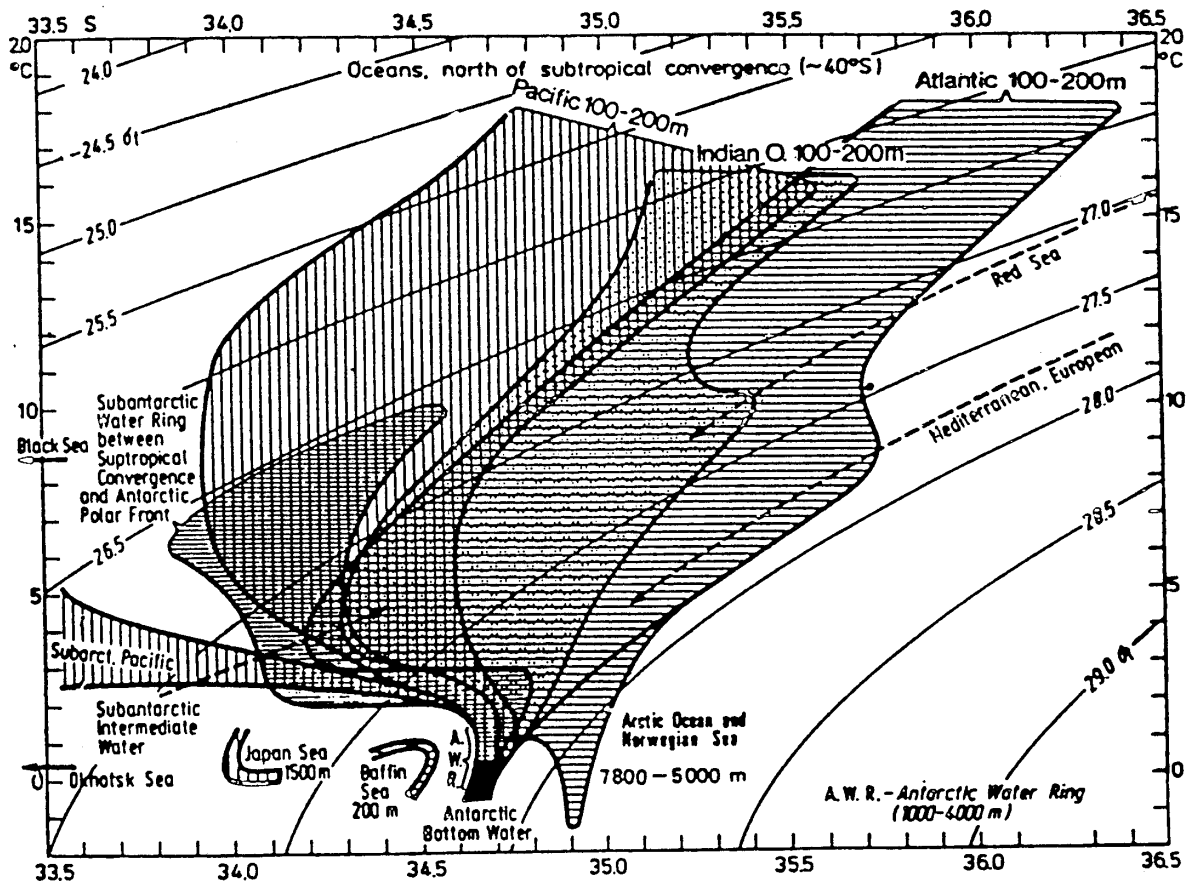

FIGURE 1.17. Temperature salinity (T-S) diagram for waters of the ocean.

The cross lines are density values. Densities increase to the lower right (higher S). Some T-S regions are unique to one water type, others are not. Surface waters (< 100m) are typically omitted. The inflections in the curves reflect the inputs of water from different sources.

How do we measure density? Normally we don't. We calculate it. The density dependence of seawater on salinity, temperature and pressure has been determined and formulated, and equations describing this relation can be used. The density of seawater is a function of temperature, pressure and salinity and is a fundamental oceanographic property. The average density (ρ) of seawater is near 1.025gm cm^{-3}. The significant part of this number is generally in and beyond the third decimal. Thus, the convention is to report density as the function $\sigma_{s,t,p} = (\rho_{s,t,p} - 1) \times 1000$.
Thus, a density of $\rho_{s,t,p} = 1.02544$ gm cm^{-3} becomes $\sigma_{s,t,p} = 25.44$.

When considering the stability of a water column, it is convenient to be able to calculate the density of a water parcel relative to its surroundings from consideration only of its temperature and salinity. In order to remove the effect of pressure on density, a parameter

9

called σ_t is calculated. This is the density of a parcel of water after it has been brought from the *in situ* depth to one atmosphere. The calculation of σ_t neglects adiabatic effects, the changes in temperature due to pressure changes. As a water parcel moves upward or downward in the ocean, its temperature varies slightly with compression. Although the compressibility of water is less than that of steel, increasing hydrostatic pressure causes an increase in the temperature of the water. This adiabatic change in temperature can be calculated from the Kelvin equation:

$$\Delta T = (T\alpha/Kc_p)g\Delta h$$

Where T = the absolute temperature (°K)
α = coefficient of thermal expansion
g = acceleration due to gravity
Δh = vertical displacement in decibars
c_p = specific heat at constant pressure
K = the mechanical equivalent of heat

Potential temperature (θ) is defined as $T-\Delta T$ where T is the temperature *in situ* and ΔT is the adiabatic temperature change caused by lifting the parcel without exchange of heat from *in situ* pressure to one atmosphere. From the potential temperature, we can calculate potential density. This is the density a parcel of water would have if brought to the sea surface adiabatically (e.g. takes into account the changes in temperature due to decompression). The potential density ($\rho_{s,\theta,0}$) is routinely reported as
$\sigma_\theta = (\rho_{s,\theta,0} - 1) \times 10^3$

The unit most commonly used to express pressure in the ocean is the decibar, which is defined as: 1 decibar = 1/10 bar = 10^5 dynes cm^{-2}
A bar is approximately equal to one atmosphere and common practice is to neglect atmospheric pressure. The decibar is a convenient unit because hydrostatic pressure increases by about one decibar per meter (or ~ 1 atmosphere every 10 meters).

Recent data from Millero *et al*. (1976) and Poisson *et al*. (1980) determined the density data with a precision of about 3 ppm. These two data sets have been combined to produce an internationally accepted one atmosphere equation of state of seawater (Millero and Poisson, 1981). The form of the equation of state is:
$$(\rho - \rho_o) = AS + BS^{3/2} + CS^2$$

Where A, B, and C are functions of temperature (°C) and S is salinity. The coefficients for the combined data are:

$A = 8.24493 \times 10^{-1} - 4.0899 \times 10^{-3} t + 7.6438 \times 10^{-5} t^2 - 8.2467 \times 10^{-7} t^3 + 5.3875 \times 10^{-9} t^4$
$B = -5.72466 \times 10^{-3} + 1.0227 \times 10^{-4} t - 1.6546 \times 10^{-6} t^2$
$C = 4.8314 \times 10^{-4}$

The absolute densities can be calculated using the value for pure water from Bigg (1967).
$\rho_o (\text{kg m}^{-3}) = 999.842594 + 6.793952 \times 10^{-2} t - 9.095290 \times 10^{-3} t^2 + 1.001685 \times 10^{-4} t^3$
$\qquad - 1.120083 \times 10^{-6} t^4 + 6.536332 \times 10^{-9} t^5$

Ocean Circulation

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 Kuroshio have a profound effect on the sea surface 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 weakening of the trade winds 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 (Wallace et al, 1998).
3. Anthropogenic pollutants including isotopes from nuclear tests contaminate the surface of the ocean. These components are slowly being mixed through the ocean and knowledge of ocean circulation is needed to determine their distribution (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 ($\sim$3.2 Gt C yr^{-1}) is less than the sum of anthropogenic emissions and deforestation ($\sim$ 7.0 Gt C yr^{-1}). Some of the CO_2 has gone into the ocean ($\sim$2 Gt C yr^{-1}). Ocean circulation is crucial for understanding this process and for predicting the effect CO_2 concentration in the future.

In this brief overview we discuss abyssal circulation, thermocline circulation and surface currents. For more thorough discussion of these and other aspects of physical oceanography take Steve Monismith's class next fall.

Abyssal Circulation
The circulation of the deep ocean below the thermocline is referred to as abyssal circulation. The currents are slow ($\sim$ 0.1 m/sec) and difficult to measure, but the pattern of circulation can be clearly seen in the properties of the abyssal water masses (temperature and salinity). 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.

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 the figure below. Warm and salty surface currents in the western North Atlantic (e.g. the Gulf Stream) transport heat to the Norwegian-Greenland Seas where the heat is transferred to the atmosphere. The cooling increases the density of seawater resulting in formation of cold and salty water in the North Atlantic. This water sinks to depth and forms the North Atlantic Deep Water (NADW). The NADW travels south through the 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 (due to sea ice formation and cooling) 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 NADW (~47%) and Antarctic margin (~53%) origin. This current 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). This deep water mass enters the Pacific in the southwest corner and flows north along the western boundary of the Tonga Trench.
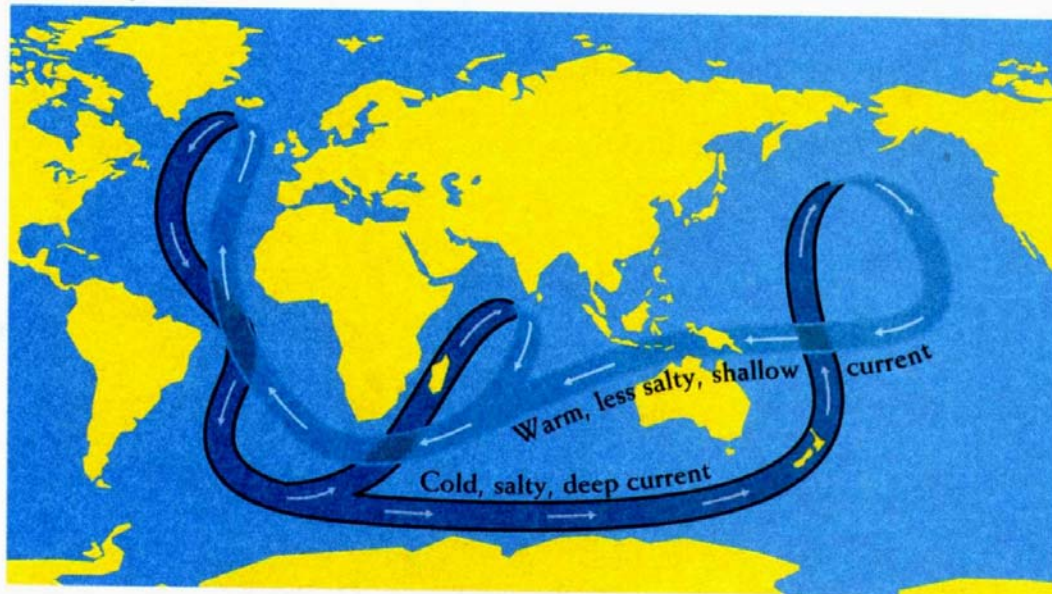
The large-scale salt transport system that operates in today's oceans. Salty, deep water formed in the North Atlantic flows down the length of the Atlantic, around Africa, through the Indian Ocean, and finally northward in the deep Pacific Ocean. This water upwells in the North Pacific, eventually working its way near the ocean surface back to the Atlantic.

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 a return flow of surface water from the Pacific to the Atlantic. There are two main paths of this return flow, which amounts to about 19Sv (1 Sverdrup = 1 million cubic meters per second). Some water passes through the Indonesian Archipelago, the Indian Ocean and around the tip of South Africa via the Agulhas Current (Gordon, 1985). Some water enters the South Atlantic via the Drake Passage. Finally there is a small transport (about 1Sv) from the Pacific to the Atlantic through the Bering Strait.
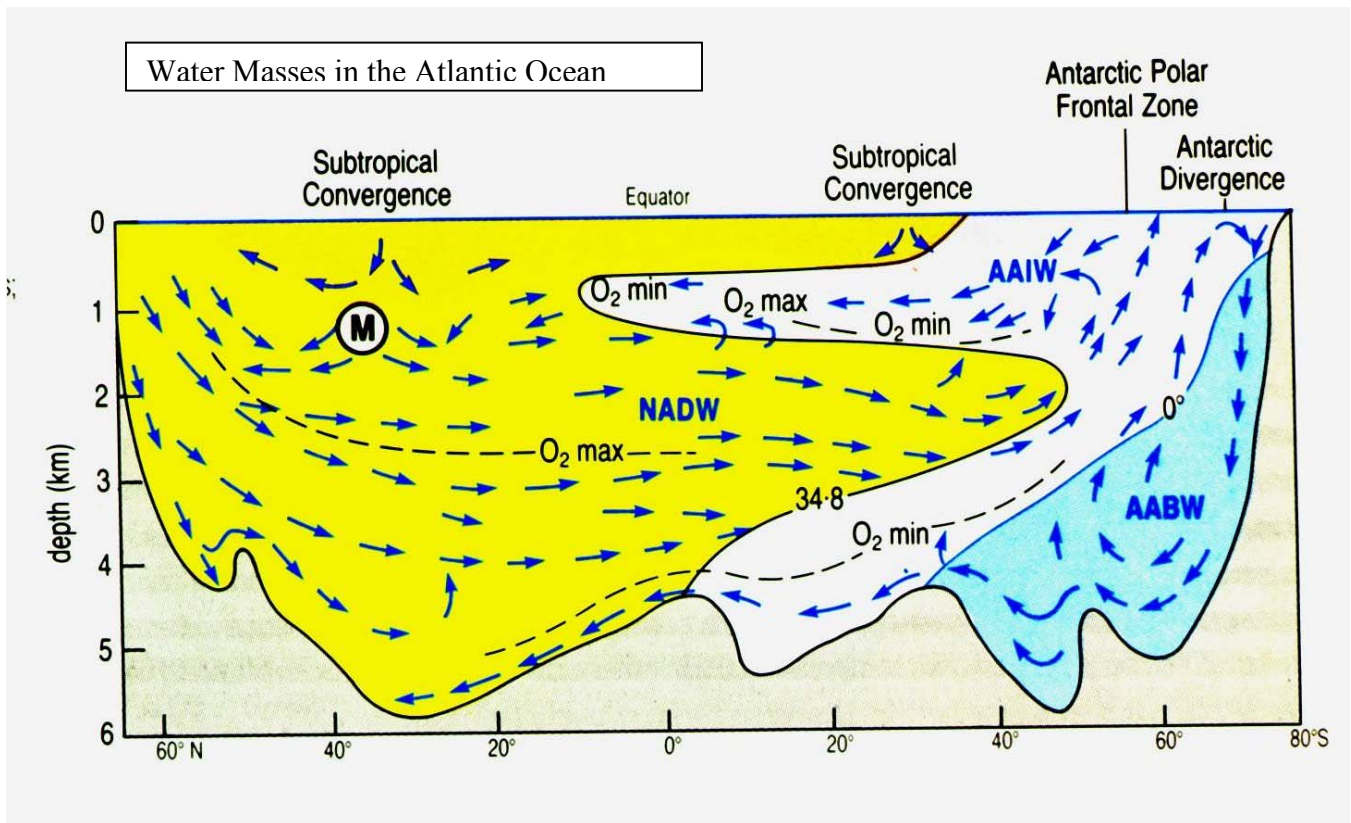
The salt budget for the Atlantic, which is determined in part by the flux of fresh water through the atmosphere, drives the conveyor belt and can explain how it has varied in the past. At present there appears to be a net water vapor loss of about 0.32Sv (greater than the flow of the Amazon) from the Atlantic to the Pacific. The NADW transports about 16.3Sv of water with a salinity of 34.91. This is produced from 15Sv of Gulf Stream water with a salinity of 35.8, 1Sv of transport from the Bering Straits with S = 32 and a net excess of river inflow and rainfall over evaporation of about 0.3Sv (Zauker and

Broecker, 1992). It is easy to show that small changes in the freshwater budget can have a significant impact on the salt budget and thus circulation. For example, if the excess of precipitation plus runoff over evaporation increased by 50% to 0.45Sv, 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 ºC and the conveyor would have to more than double its flow to restore the salt balance (Broecker, 1997).

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. Using ^{14}C distribution in deep water 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 ~ 4.6 million years ago (Keigwin, 1982; Maier-Reimer et al., 1990). 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, helping to 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 seem 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 iceberg input, mainly from Canada (Bond et al., 1992). These icebergs transport terrigenous debris across the North Atlantic. When they melt they deposit a layer of 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 slowing or shutting down the present mode of the conveyor belt. The record 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, 1995). At the time of these events 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 time scale of a few thousand years. These so-called Dansgaard-Oeschger cycles are characterized by abrupt changes in temperature, dust content, ice accumulation rate and methane concentration in the atmosphere. 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, iceberg 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 thousand year time frame. These cyclic events appear to have continued in the Holocene, although with significantly muted amplitudes (Alley et al., 1997).
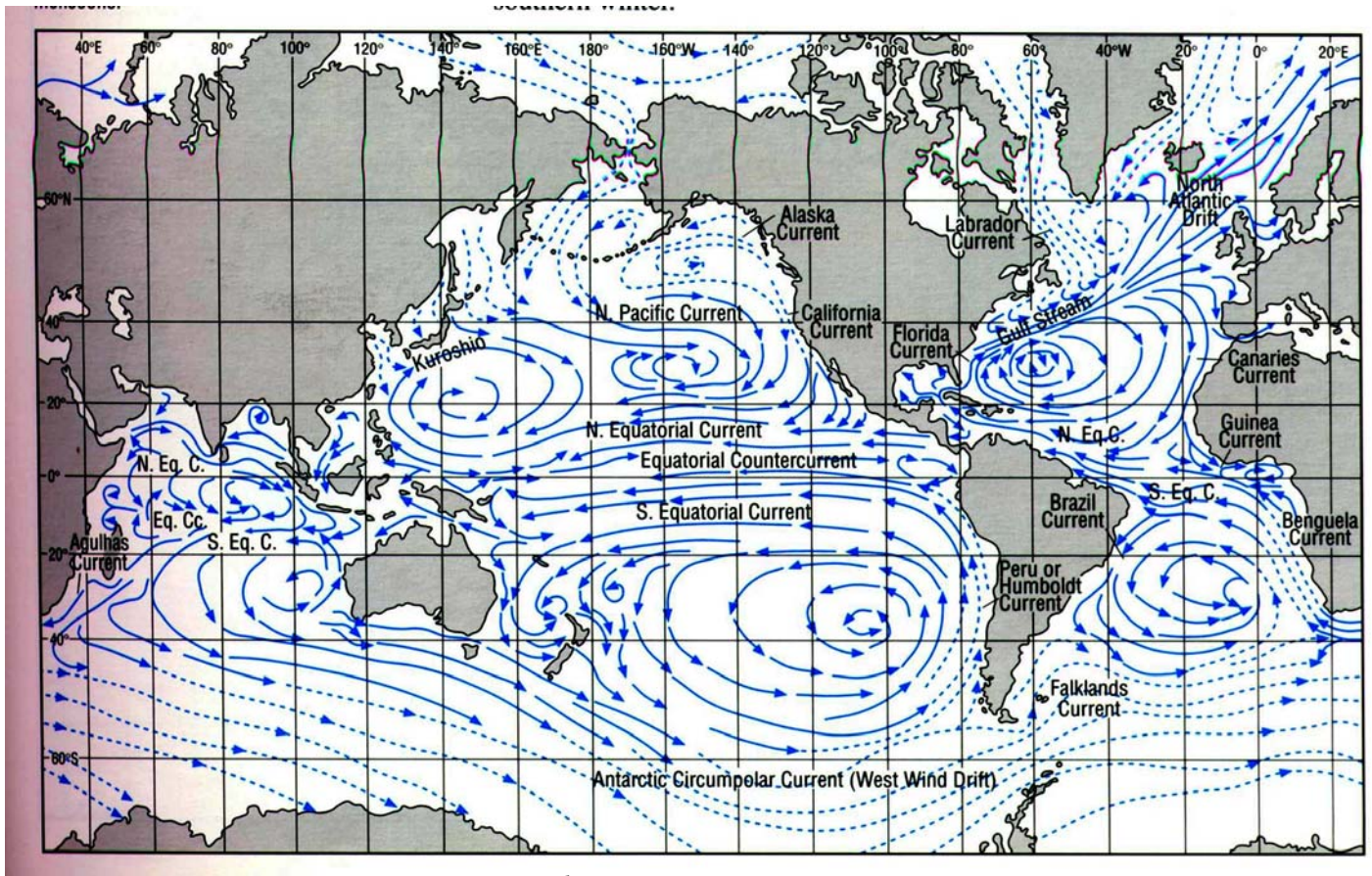
Surface Currents

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 (down-welling) 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. The wind blowing across the sea surface 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 an Ekman spiral. Ekman transport, changes in sea surface topography and the Coriolis force combine to form geostrophic currents. In 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 sea surface "topographic high". As a result of the elevated sea surface, 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 sea surface 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 sea surface and clockwise. 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 counterclockwise gyres in the southern hemisphere. 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. Where water diverges there is upwelling and where it converges, downwelling (Thurman, 1990). This circulation has important consequences for chemical oceanography because regions of upwelling (near the continental margins and at the equator) are locations of high nutrient content in the surface waters and locations of downwelling (the subtropical gyres) are locations of very low nutrient concentration.

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 Sv = 1 x 10^6 m^3s^{-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.

Oceanic Surface Currents　　　　　**Upwelling in the Ocean**
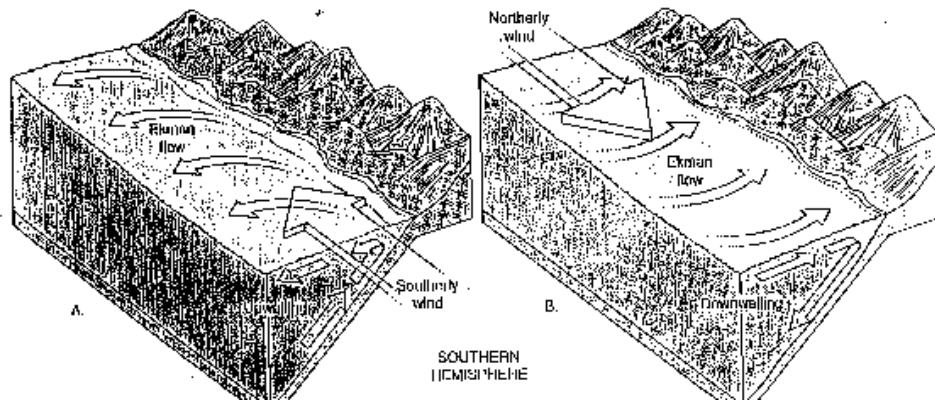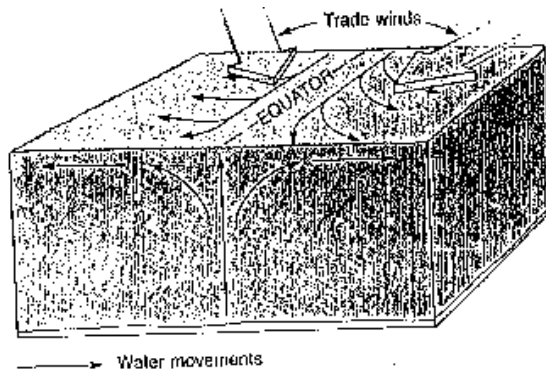
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