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"Thermohaline Forcing & Heat Fluxes"

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Thermohaline Forcing and Heat fluxes

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1 Sea water

Sea water is a complex mixture : about 96.5 % water, 3.5 % salt, minute amounts of living things, and usually an equal amount of dust and dissolved organic matter.

Water's unusual properties arise from its asymmetrical molecular structure: such molecules are called **polar molecules**. The two hydrogen atoms form weak hydrogen bonds with the oxygens of adjacent water molecules: these bonds are responsible for water's ability to store large amounts of heat energy with relatively small temperature changes.

Heating of the ocean surface by the sun provides enough energy to break hydrogen bonds. Energy absorbed through evaporating water (breaking hydrogen bonds) and released through condensing water (reforming hydrogen bonds) is important in exchanging heat between ocean and atmosphere.

Below are shown some anomalous physical properties of water and their effects on the ocean:

- Heat capacity
  - Highest of all solids and liquids except liquid ammonia
  - Prevents extreme ranges in ocean temperature. Heat transfer by current is large
- Latent heat of fusion

- Highest except ammonia
- Acts as thermostat at freezing point owing to uptake or release of latent heat

- Latent heat of evaporation
  - Highest of all substances
  - Extremely important in heat and water transfer to atmosphere

- Thermal expansion
  - Temperature of maximum density decreases with increasing salinity. For pure water, it is at 4 $^{\circ}C$
  - Fresh water and dilute seawater reach maximum density at temperatures above freezing point

- Dissolving power
  - Dissolves more substances and in greater quantities than any other liquid
  - On the ocean there are many kind of dissolved substances

- Transparency
  - Relatively great
  - Absorption of radiant energy is large in infrared and ultraviolet, in visible portion of energy spectrum is relatively little.

1.1 Temperature

Temperature affects water's internal structure and its properties. Much of the heat absorbed by water is used to change its internal structure. Thus after absorbing a given amount of heat, water has a lower rise in temperature than do other substances; that is water has a high heat capacity.

The large amount of water on the earth's surface prevents wide variations in surface temperatures. Much of the incoming solar radiation reaching the earth's surface goes into evaporating water and melting ice. In the

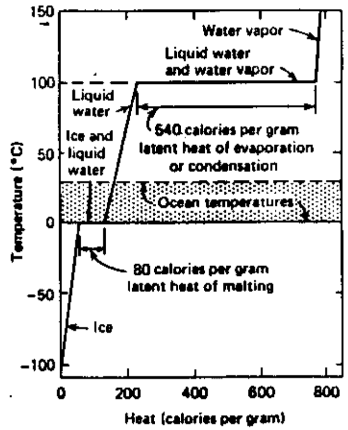

Figure 1: Temperature changes when heat is added to or removed from

ocean, water temperatures rarely exceed 30 $^{\circ}C$ or go below -2 $^{\circ}C$. Let us see what happens to water during these changes and how it affects the earth's heat balance.

First, we define a measure of heat, the **calorie**, as the amount of heat required to raise the temperature of 1 gram of liquid water by 1 $^{\circ}C$. Breaking bonds in ice and liquid water requires energy, usually heat. Conversely, heat is released when the bonds re-form. To illustrate this process, consider what happens when 1 gram of ice just below the freezing point is slowly heated. As we add heat to the ice, its temperature increases about 2 $^{\circ}C$ for each calorie of heat we add.

When the ice reaches its melting point (0 $^{\circ}C$), the temperature remains constant. At that time the heat added is going into breaking hydrogen and van der Waals bonds. After we have added about 80 calories, enough hydrogen bonds have been broken and the last bit of ice melts, leaving only liquid water.

As we continue to supply heat, the water temperature rises but at a lower rate, 1 $^{\circ}C$ per calorie of added heat. This rate of temperature change remains nearly constant between 0 $^{\circ}C$ and 100 $^{\circ}C$. At 100 $^{\circ}C$ (the boiling point of liquid water), water vapor begins to form and the temperature remains constant while both water vapor and liquid water are present.

After we have added about 540 calories, all remaining hydrogen bonds are broken, so each water molecule exists alone , in other words, as a gas. If we capture the gas and continue to heat it, we find that the temperature rises again at 100 $^{\circ}C$ per calorie of heat added.

If we started with water vapor and cooled it, we would reverse this process and gain heat each time there was a change in state. Heat added to the water is taken up in two ways. One is **sensible heat**, heat that we detect either through touch or with thermometers. This temperature change results from the increased vibration of molecules or motions in the gas state.

The other form is called **latent heat**, which is the energy necessary to break bonds in water. As we have seen, at the melting point and the boiling point of water there was no change in temperature as long as two states of matter coexisted. As discussed above, added energy went into breaking bonds in the disappearing phase. This is called latent heat because we get back all the heat when the process is reversed. Condensing water vapor to form liquid water releases 540 calories per gram at 100 $^{\circ}C$; freezing water at 0 $^{\circ}C$ releases 80 calories per gram. The difference between the latent heats of melting and evaporation arises from the fact that only a small fraction of the hydrogen bonds are broken when ice melts. All hydrogen bonds are broken when water evaporates.

Although water freezes at 0 $^{\circ}C$ and boils at 100 $^{\circ}C$, water molecules can go from vapor to solid or liquid at other temperatures. The processes involved are similar to those described earlier, but the amount of heat involved is different. For example, the latent heat of evaporation changes as follows.

Temperature	Latent heat of evaporation
0 $^{\circ}C$	595 *cal*
20 $^{\circ}C$	585 *cal*
100 $^{\circ}C$	539 *cal*

More energy is required to remove a water molecule from liquid water at 0 or 20 $^{\circ}C$ than 100 $^{\circ}C$. This is important in the ocean because most water evaporates from the ocean surface at temperatures around 18 to

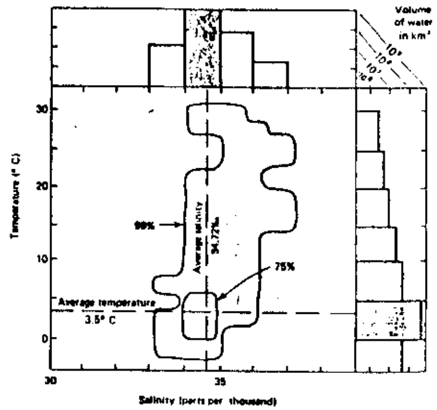

Figure 2: Temperature and salinity of 99 % of the ocean water

20 $^{\circ}C$. If water had to reach 100 $^{\circ}C$ before evaporating, we would have no water vapor in the atmosphere.

1.2 Salinity

The amount of salt in a volume of seawater varies from place to place due to local additions or removals of water. To indicate the amount of salt in a seawater sample, oceanographers use **salinity**, wich is defined as the amount of salt (in grams) dissolved in one Kilogram of seawater. Salinity is thus expressed in parts per thousand ($^{o}/_{oo}$).

Salinity is determined by measuring seawater's **electrical conductivity**, its ability to conduct an electrical current, which is controlled by movements of ions through the water.

1.3 Density

Whether a substance sinks or floats in a liquid is determined primarily by its **density** (mass per unit volume) in relation to its surrounding. The density maximum in pure water (4 $^{\circ}C$) is important in lakes. In the ocean the presence of salt eliminates the density maximum. The temperatures of both maximum density and freezing decrease as salinity increases.

But the temperature of maximum density drops more rapidly than the freezing temperature. Thus seawater of normal salinity freezes before

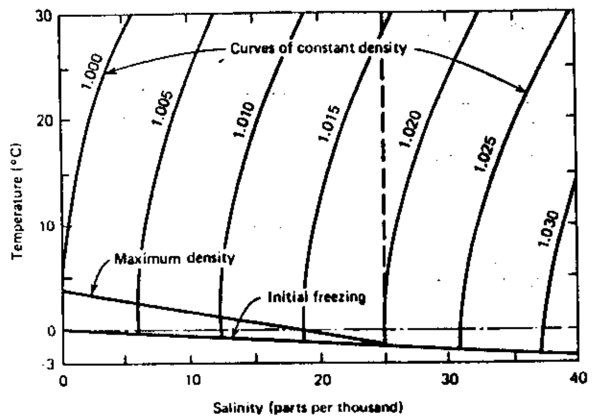

Figure 3: Variation in seawater density as function of temperature and salinity

it reaches the temperature of maximum density.

The density of seawater is determined primarily by temperature and salinity. (Pressure is important only in the deepest parts of the ocean basins and will be ignored in this discussion.) Decreases in temperature and increases in salinity cause increased density. Reading along the top line of the graph, you will note that an increase in density from 1.015 to 1.020 grams per cubic centimeter can be achieved by increasing salinity from 26 to 33 $^o/_{oo}$ (at a constant 30 ^oC). By reading along the dashed line, you will find that a comparable density change results from increasing the temperature from approximately 5 to 29 ^oC (at a constant 25 $^o/_{oo}$). Salinity at the ocean surface does not vary greatly from about 35 $^o/_{oo}$. Thus density changes in the upper layers of the ocean are mainly caused by changes in water temperatures.

The ocean is normally stably stratified, with the densest waters near the ocean bottom and the least dense waters nearest the surface. When unstable density distributions occur, they are usually modified rather quickly by convective overturning, whereby the denser waters sink to their appropriate levels. We shall learn more about such situations when we discuss the formation of dense water masses later.

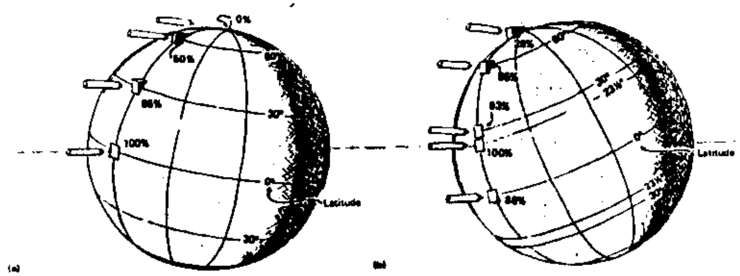

Figure 4: Incoming solar radiation varies over the earth's surface: (a) Northern Hemisphere at the equinoxes. (b) Summer solstice (June 21)

2 Heat budgets

Budgets are useful to indicate where materials or energy comes from (sources) and where they go (sinks). We start with the earth's heat budget. The earth is exposed to the sun's radiation from a distance of 149 million kilometers. The top of the atmosphere receives about 2 calories per square centimeter per minute. (2 $cal\ cm^{-2}\ min^{-1}$) Despite the large amount of heat received by the earth from the sun, the earth's surface temperatures over the past few centuries (for which we have historical records) have been nearly constant. Therefore, we assume that the earth radiates back to space each year as much energy as it receives from the sun. The figure below show the major interactions with incoming solar radiation and loss of heat from the earth (annual average heat budget). About one-third of the solar radiation striking the top of the atmosphere is reflected back to space by clouds. On the average, about half of the sun's radiation striking the top of the atmosphere reaches the earth's surface. The atmosphere is heated in the lowermost few thousand meters primarily through the latent heat from evaporated water, principally from the ocean. The heat that originally went into evaporating water is released to the atmosphere when water vapor condenses. Relatively little heat is transferred directly by sensible heat, where a volume of warm air flows into a region and an equal volume of cold air flows out. Averaged over a year, the maximum amount

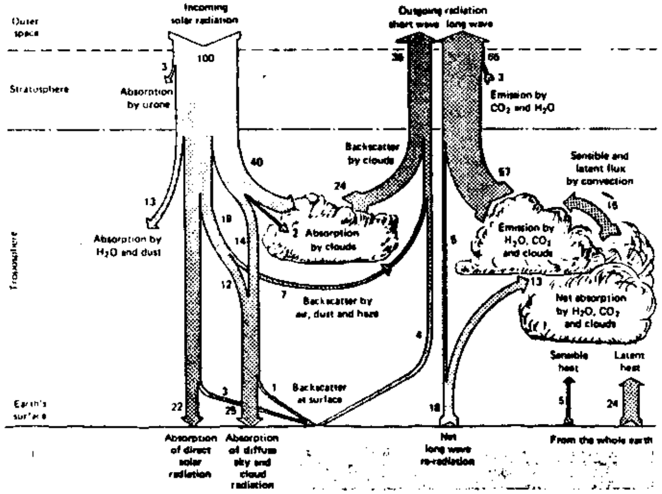

Figure 5: Average heat budget interactions

of incoming solar energy is received in low latitudes. The atmosphere , however, radiates energy back to space at a nearly constant rate. As a result, the earth receives most of its heat in low latitudes; large inputs of solar energy into tropical regions cause high ocean surface temperature here.

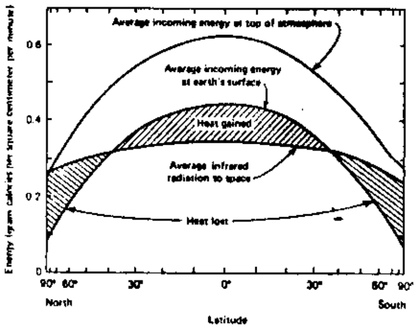

Figure 6: Areas of heat loss and heat gain on the earth's surface

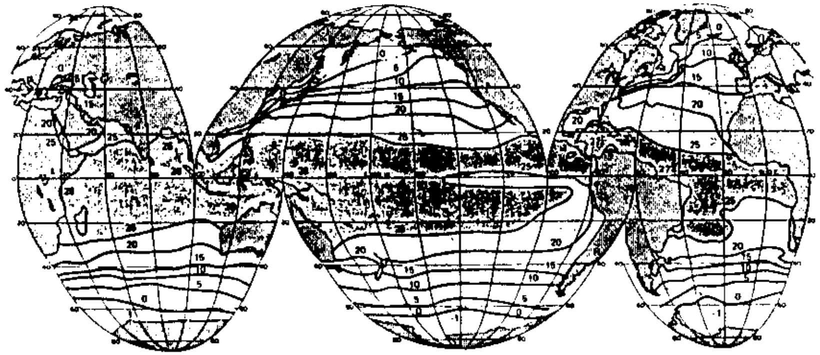

Figure 7: Winter sea surface temperature

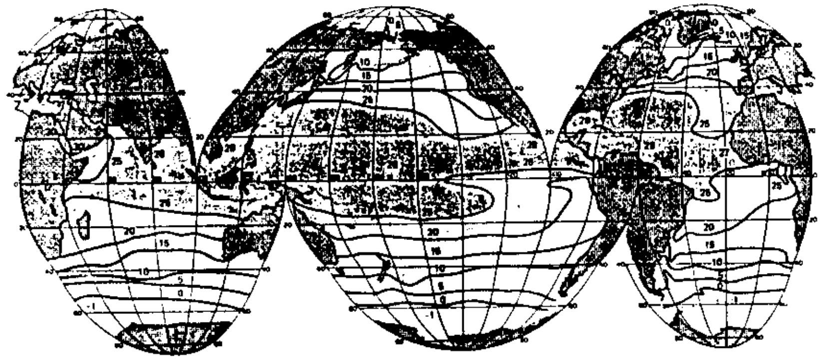

Figure 8: Summer sea surface temperature

3 Sea surface temperature

Bands of equal ocean surface temperatures are oriented east-west and centered on the equator. Temperatures are higest along equator because of the warming of the earth in the tropics. They gradually become cooler toward the poles. There are obvious complications because currents transport cold water toward the equator and warm waters toward the poles.

Ocean temperatures change with the seasons, due to variations in the amounts of incoming solar radiation. In equatorial regions, water and air temperature change little seasonally. The small changes in polar ocean

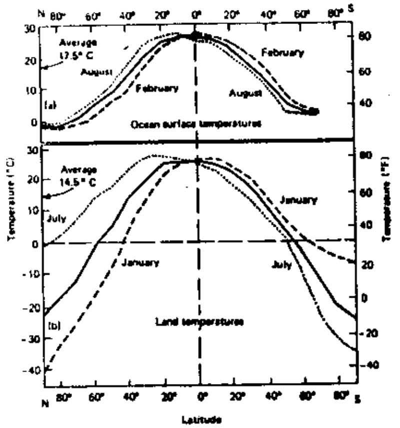

Figure 9: Average and range of temperatures at the earth's surface

water temperatures are due to the year-round presence of ice. Differences in surface ocean temperatures between February and August are greatest in midlatitudes. By contrast, on land the largest seasonal temperature differences occour in the high latitudes, where the presence of sea ice prevents the ocean from moderating atmospheric temperatures as it does elsewhere.

4 Sea surface salinity

Sea surface salinities are distributed quite differently than are water temperature. The highest salinity waters occur in the centers of the ocean basin. Lowest salinities occur near continents, where the coastal ocean receives large river discharges, and in the high latitudes. Salinity differences are caused by regional variations in evaporation and precipitation.Evaporation from the sea surface is greatest in subtropical areas, where it is favored by clear skies and dry winds. The highest surface water salinities occur in the centers of ocean basins, where there is no diluition by river disharges. The sides of ocean basins where dry winds blow off the continents are also areas of high evaporation. Discharges of rivers often dilute the coastal waters, obscuring the effects of high evaporation. The relatively low salinities of the high latitudes are due, in part, to low rates

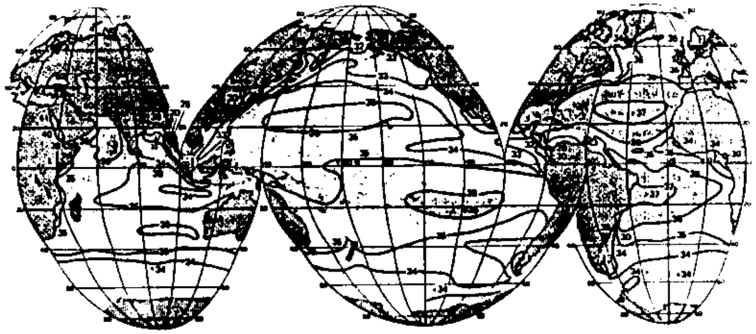

Figure 10: Sea surface salinity

of evaporation there.

Precipitation is highest near the equator, equatorial surface waters are diluted by the heavy rainfall resulting from the condensation of warm, moist air rising in the equatorial atmospheric circulation. River runoff from the continents dilutes coastal ocean waters, causing them to have relatively low salinities.

In summary, the highest salinity surface waters occur in areas of excess evaporation, either in subtropical central regions of the ocean basin or in landlocked seas in arid regions. The lowest salinities occur in areas where precipitation exceeds evaporation, primarily in coastal or equatorial regions.

5 Oceanic budgets of heat, water, and salt

Water and heat budgets are intimately related. Heat is removed from the ocean by evaporation of water from the ocean surface. This heat is released to the atmosphere when water vapor rises, cools, condenses, and falls back to earth as rain or snow (latent heat).

We know that the amount of water on the earth's surface has remained constant for billions of years: thus we can make a budget to keep track of where the water goes, how it moves, and where it is stored.

The ocean is the primary reservoir for water on the earth's surface:

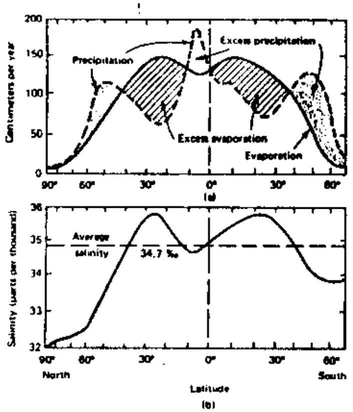

Figure 11: (a) relation between evaporation and precipitation (b) surface water salinities

One convenient way to visualize the earth's water budget.

Area	Precipitation	Evaporation
Ocean	88 cm/y	97 cm/y
Continent	67 cm/y	47 cm/y
Entire Earth	82 cm/y	82 cm/y

A complete water budget for a region can be written as follows:

$$E = P + R + T$$

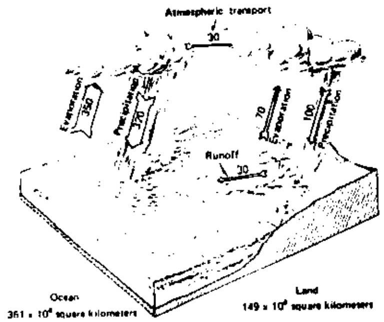

Figure 12: Hydrologic cicle

where E = *evaporation*, P = *precipitation*, R = *river runoff*, T = *current transport*.

If the ocean region under consideration is large, the amount of river runoff compared with the total amount of seawater is small. The currents in most regions simply move water around , thus both terms can be ignored. So the water budget for a large ocean region simplifies to

$$Evaporation = Precipitation.$$

It is possible to make a heat budget because we know that earth loses as much heat as it gains each year. The full heat budget can be written as follows:

$$H_{SR} = H_E + H_R + H_A + H_C$$

where H_{SR} = Heat from solar radiation, H_E = Heat lost by evaporation, H_R = Heat lost by radiation, H_A = Heat lost by heating atmosphere, H_C = Heat lost by currents.

Again, if the ocean region is large (say an entire basin), the amount of heat gained or lost by current transport can usually be ignored. Thus the heat budget says that the heat from the sun's radiation is lost through evaporation, radiation back to space, and to direct heating of the overlying atmosphere.

Areas of relatively high surface salinity supply heat to the atmosphere in the form of water vapor, which is transported by winds to equatorial or subpolar regions. There water vapor condenses, forming either rain or snow, and releases the heat to the atmosphere. Thus the poleward transport of heat and water is reflected in the relatively low surface water salinities in high latitudes. Regional salinity differences of surface ocean waters are manifestations of the worldwide heat transport system of the ocean and atmosphere.

6 Depth zones

There are three major depth zones in the ocean : surface, pycnocline (the density boundary between the surface and deep ocean), and deep

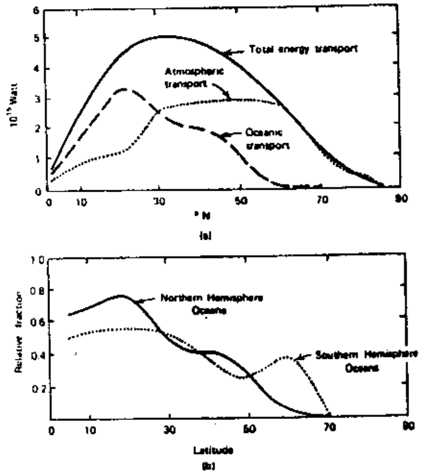

Figure 13: Energy transport as function of latitude

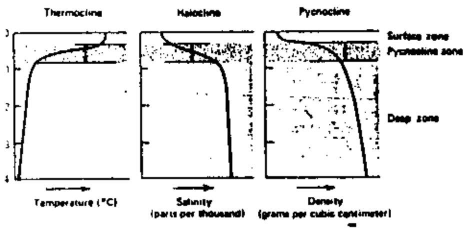

Figure 14: Thermocline, halocline, picnocline

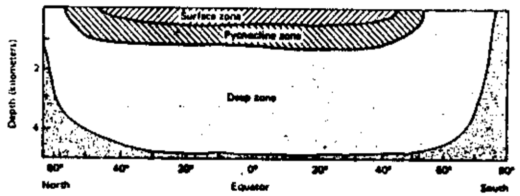

Figure 15: Schematic representation of the density structure of the ocean

zones. The **surface zone** is in contact with the atmosphere. It undergoes substantial changes because of seasonal variability in local water budgets (evaporation, precipitation) and heat budgets (incoming solar radiation, evaporation). The surface zone sometimes called the **mixed layer** is typically 100 - 500 meters thick and constitutes about 2 % of the ocean's volumes.

In the **pycnocline zone**, water density changes markedly with increasing depth and thus forms a layer of much greater stability than the overlying surface water. Formation of a pycnocline zone results from marked changes with depth in either salinity or temperature. There the pycnocline coincides with the **thermocline**, or zone of sharp temperature change. Thermoclines are especially prominent in open-ocean areas, where the surface layers are strongly heated during much of the year. At higher latitude there is less surface heating by incoming solar radiation. There the lowered surface salinity caused by abundant precipitation causes a **halocline**, a marked change in salinity with depth. The pycnocline partially isolates the deep ocean from surface processes because low density surface waters cannot move downward through it. Deep ocean waters move slowly upward through the pycnocline as part of the worldwide deep ocean circulation.

The **deep zone** contains about 80 % of the ocean's waters and lies at depth of about 1000 meters , except at high latitude, where deep water are in contact with the atmosphere. The schematic representation of the density structure of the ocean show that the pycnocline zone isolates the deep ocean from contact with the atmosphere in the equatorial and midlat-

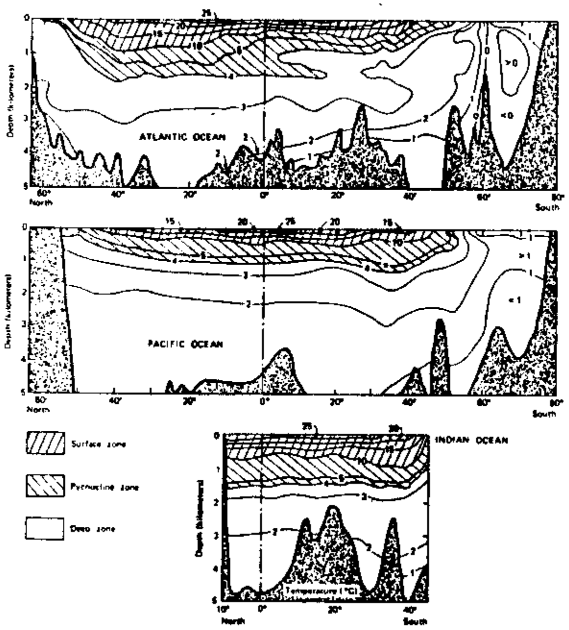

Figure 16: Vertical distribution of temperature in all three oceans

itude areas. Only near the North and the South poles does the deep ocean have any contact with the atmosphere. The figures of the vertical distribution of temperature and salinity in all three ocean basins shown a strong similarities with the schematic representation of the density structure we previously considered. We can also note the complexities of temperature and salinity distribution evident on even such simplified diagrams.

The base of the surface zone corresponds approximately to the 10 °C isotherme contour on the vertical section. The base of the pycnocline can be taken as the 4 °C isotherm. Note the strong similarities between the distribution in the three ocean basins and note that there are differences between hemispheres even within the same ocean basin. Water masses are formed in a few ocean areas, usually partially isolated seas or on continental shelves. If a mass of surface water become more dense that the water below, it sinks below the surface.

Temperature and salinity of subsurface waters are used to study

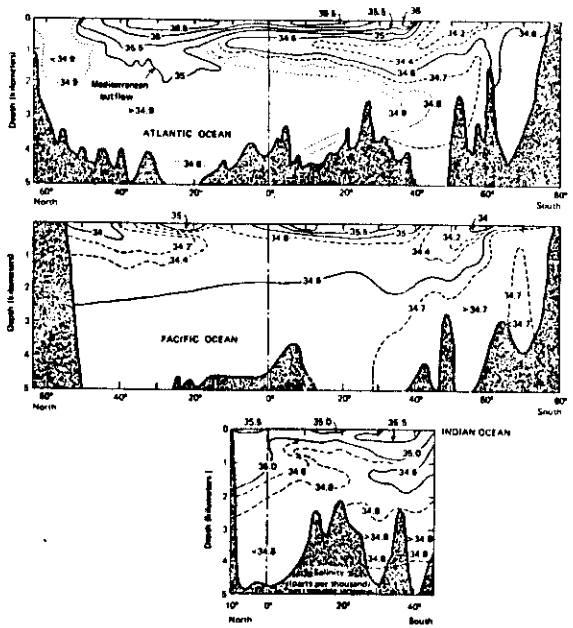

Figure 17: Vertical distribution of salinity in all three oceans

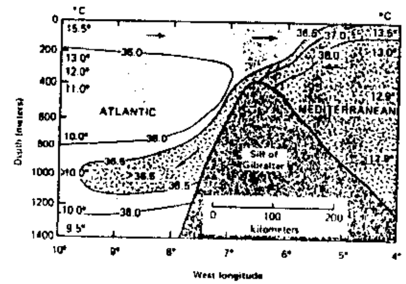

Figure 18: Salinity distribution through the Strait of Gibraltar

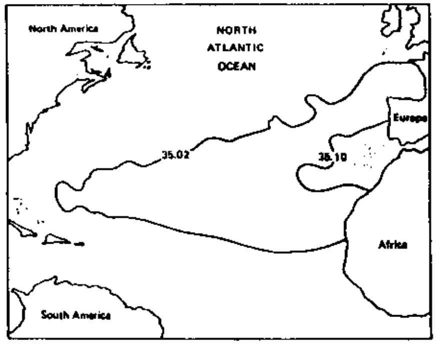

Figure 19: Horizontal distribution of salinity at depths around 1000 m

movements of well defined subsurface water as they spread out from a source region. An example is the warm, saline water formed in the Mediterranean sea that flows out through the Strait of Gibraltar, near the bottom and spreads laterally at mid depth (about 1000 m) in the North Atlantic ocean.

7 Thermohaline Circulation

The general pattern of subsurface water movements is determined by the density distributions of oceanic waters. Recall that the depth of a particular water mass is determined by its density relative to the densities of waters around it and that when a water mass reaches its appropriate density level, it tends to spread out, forming a thin layer. This is a reflection of the fact that less energy is required to move a parcel of water along a surface of constant density than across it. Remember also that the ocean is primarily horizontally layered, as shown by the structure of the surface; pycnocline, and deep ocean zones. Thus surfaces of constant density are essentially horizontal. Consequently, subsurface water motions are primarily horizontal along surfaces of constant density.

There are areas in the ocean of vertical water movements, primarily

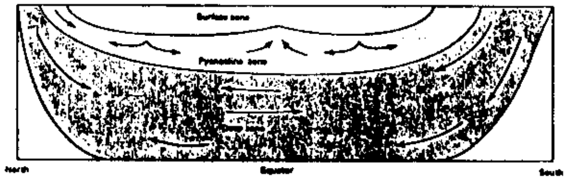

Figure 20: Schematic representation of the thermohaline circulation

in the high latitudes where dense water masses are formed. There is some evidence of sinking due to winter cooling in the Labrador Sea South of Greenland but it is probably very localized in both space and time. In the South Atlantic, the main source is probably the Weddell Sea where sinking results from density increase due to freezing out of ice. The process in both North and South are thermohaline and, of corse, seasonal. The resulting massive vertical water movements control water temperatures and salinities throughout the deep ocean and drive the major currents along the bottom and in the mid-depths of the ocean. These density-driven currents are known as the thermohaline circulation because of the important role played by temperature and salinity in controlling seawater density. These currents transport cold waters from polar regions. The waters involved slowly return to the surface (after many hundreds of years), some through the pycnocline throughout the ocean, and others by upwelling along the equator and in coastal regions.

In addition to the deep water formation due to winter cooling, there are also contributions of thermohaline origin at mid-depth from the Mediterranean to the Atlantic and from the Red Sea and Persian Gulf to the Indian Ocean. These are waters rendered dense by evaporation at the surface which then sink and flow out of the seas into the neighbouring oceans.

Stommel has put forward ideas for a model of the circulation of the deep waters. He brings in another feature of the ocean structure, that the depth of the thermocline at any locality remains substantially constant. Because in low latitudes there is a net annual inflow of heat through the surface into the water, the upper warm layer, and its boundary the ther-

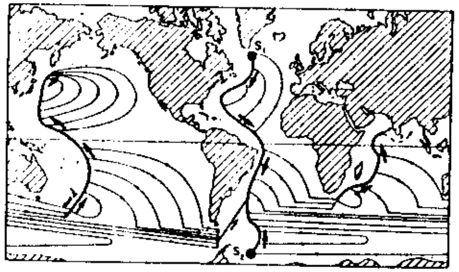

Figure 21: Model for deep ocean circulation

mocline, should deepen with time. As this deeping does not happen, some mechanism must be opposing the tendency, and Stommel suggests that this mechanism is slow upward flow of cool deep water. Continuity requires that the sinking water in the North and South Atlantic must be balanced by rising, and Stommel suggests that while the sinking is very localized, the rising is spread over most of the low and middle latitude areas of the oceans. The sinking regions (S_1,S_2) are shown feeding relatively intense western boundary currents.

Stommel makes it clear that he does not regard the above as a *theory* but as a *model* for quantitative study which might form the basis for a theory.

To produce a theory with thermohaline effects included may require taking T and S (and hence ρ) not as given by observations but as unknowns which must be solved for, in the problem, along with the velocity. If T and S are to be unknowns then we require equations for them.

8 Equations

The differential equations for temperature and salinity are

$$\frac{dT}{dt} = \kappa_T \nabla^2 T + Q_T \tag{1}$$

$$\frac{dS}{dt} = \kappa_S \nabla^2 S + Q_S \tag{2}$$

where κ_T and κ_S are molecular kinematic *diffusivities* for temperature and salt (κ_T is about 100 time κ_s), and $\nabla^2 = \frac{\partial^2}{\partial x^2} + \frac{\partial^2}{\partial y^2} + \frac{\partial^2}{\partial z^2}$, Q_T and Q_S represent a source function. For example, if solar radiation is absorbed over a significant depth range and causes heating, Q_T represents such effects. Q_S is a similar function for processes affecting salinity ; it is important only at boundaries, e.g. river inputs, effects of freezing or the difference between evaporation and precipitation.

These equations apply to an individual, small (mathematically infinitesimal) fluid element, and apply to the total instantaneous values. If we recall the Eulerian form of the total derivative and we adopt Reynold's approach of splitting the total quantities into mean and fluctuating part: $T = \overline{T} + T'$ and take the average of the equation, we have

$$\frac{\partial \overline{T}}{\partial t} + \overline{u}\frac{\partial \overline{T}}{\partial x} + \overline{v}\frac{\partial \overline{T}}{\partial y} + \overline{w}\frac{\partial \overline{T}}{\partial z} + \overline{u'\frac{\partial T'}{\partial x}} + \overline{v'\frac{\partial T'}{\partial y}} + \overline{w'\frac{\partial T'}{\partial z}} = K_T \nabla^2 \overline{T} + Q_T \tag{3}$$

The first four terms may be written as $\frac{d\overline{T}}{dt}$ for the total derivative following the *mean* flow. The next three represent the effects of turbulence on the temperature field. The diffusivity term looks the same as before except that $\overline{T}$ replaces T.

Using the continuity equation for the fluctuating velocity , we can rewrite the turbulent terms : then we have the term $\frac{\partial \overline{u'T'}}{\partial x}$. If we suppose that, by analogy with the molecular case, the turbulent fluxes are related to the mean gradients, we have:

$$\overline{u'S'} = -K_{S_x} \frac{\partial \overline{S}}{\partial x} \tag{4}$$

Now the turbulent mixing is dominated by the turbulent flow field so it is common assume that the eddy diffusivity is the same for all scalars (unlike the molecular values). Because of the static stability, K_{S_z} will be much smaller than K_{S_x} , K_{S_y} but these two should be similar. Thus we replace K_{S_z} by K_z (vertical eddy diffusivity) , and the other two by

K_H (the horizontal eddy diffusivity). Finally, neglecting the variations of the K's with space coordinates, neglecting the K_s term compared with the turbulence terms and dropping the overbar for simplicity, equations becomes :

$$\frac{dT}{dt} = K_H\left(\frac{\partial^2 T}{\partial x^2} + \frac{\partial^2 T}{\partial y^2}\right) + K_z\frac{\partial^2 T}{\partial z^2} + Q_T \tag{5}$$

$$\frac{dS}{dt} = K_H\left(\frac{\partial^2 S}{\partial x^2} + \frac{\partial^2 S}{\partial y^2}\right) + K_z\frac{\partial^2 S}{\partial z^2} + Q_S \tag{6}$$

9 Thermoclines and thermohaline circulation

Let us consider the steady state case and ignore Q_T because we are interested in the main thermocline. Then equation for T becomes

$$\vec{V}_H \nabla_H T + w\frac{\partial T}{\partial z} = K_H \nabla_H^2 T + K_z \frac{\partial^2 T}{\partial z^2} \tag{7}$$

where $\vec{V}_H$ is the horizontal velocity and ∇_H and ∇_H^2 are horizontal gradient and Laplacian respectively. Our knowledge of the deep circulation is not sufficient to allow us to drop any of the terms as being small. Stommel's conceptual model suggests that both advective terms are needed in a thermohaline circulation theory and at least the vertical diffusion term. Lateral diffusion may be well important too. It is possible to try to balance pairs of these four terms while neglecting the other two to see what solutions are possible.

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

$$\frac{\partial^2 T}{\partial z^2} = \frac{w}{K_z}\frac{\partial T}{\partial z} \tag{8}$$

Assuming $\frac{w}{K_z}$ independent of z and $T = T_d$ for $z \ll -K_z/w$, then $T = T_d + (T_0 - T_d)exp(wz/K_z)$ where T_0 is the temperature at $z = 0$ (taken to be at the bottom of the mixed layer). Adding a mixed layer on top and adjusting w/K_z one can produce a reasonable fit to observed vertical temperature profiles.

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"Thermohaline Forcing & Heat Fluxes"

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Istituto Studio Dinamica Grandi Masse
CNR
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Mediterranean Average vertical profiles of Salinity

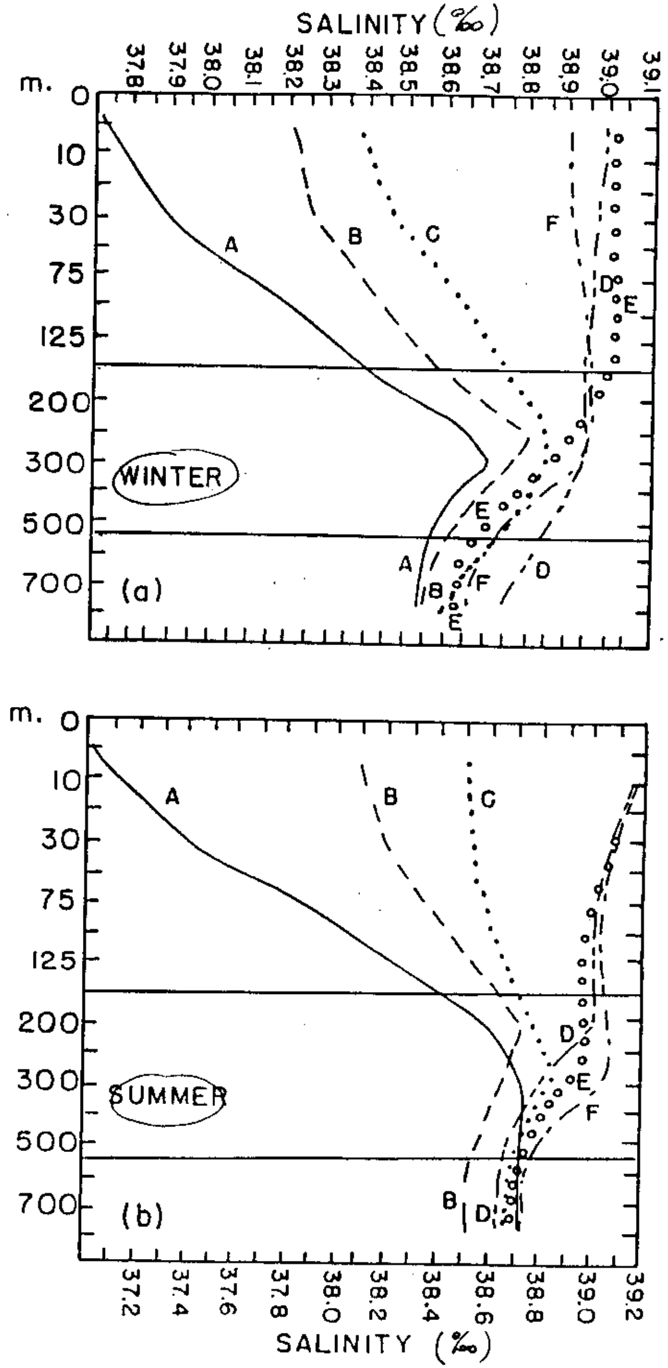

Fig 1

Mediterranean
Annual Average of latent heat in Watts/m^2

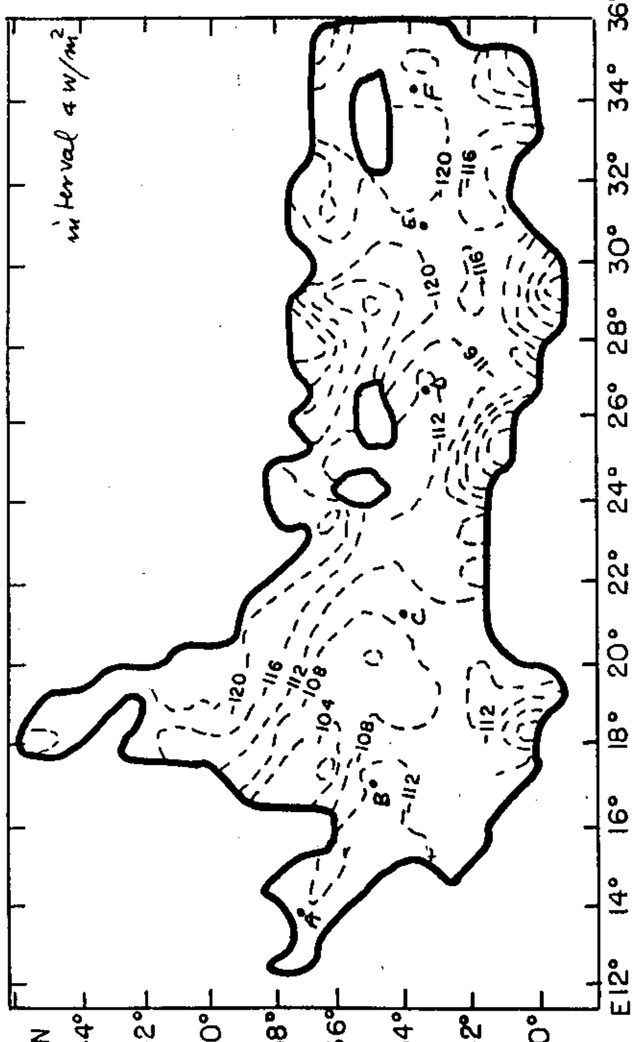

Fig. 2

Total heat budget in Watts/m^2

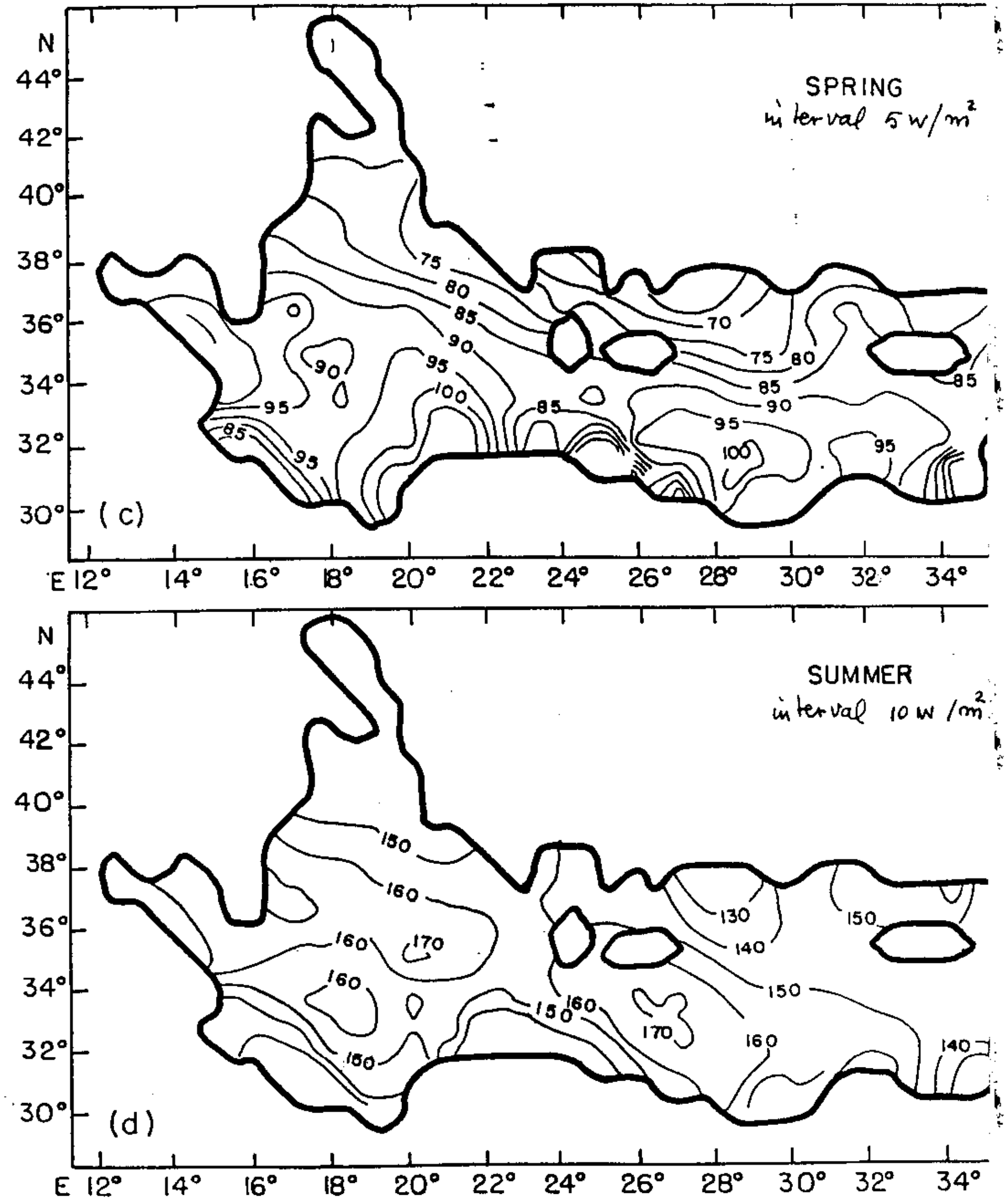

Fig 3

Total heat budget in watts/m²