

PHYSICAL OCEANOGRAPHY

by

Tor Gammelsrød

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The material is collected from different sources. The most important are:

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Segar, D.A. Introduction to Ocean Sciences. Wadsworth, 1997

Stowe, K.: Essentials of Ocean Science, John Wiley & Sons 1987

Tomczac, M. & S. Godfrey: Regional Oceanography, An Introduction. Pergamon, 1993

see also: <http://gaea.es.flinders.edu.au/~mattom/IntroOc/>

The Open University/Pergamon press:

Ocean Circulation

Waves, Tides and shallow water processes.

<http://gaea.es.flinders.edu.au/~mattom/IntroOc/>

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1. HOW EVERYTHING IS DRIVEN

The basic for all life on the earth is the sun. Also all winds and currents (except for the tides) are indirectly driven by the sun. The fact that the radiation from the sun is not evenly distributed by latitude, time (annual and diurnal) gives rise to a variety of motions and processes.

Fig.1.1 indicates the average radiation balance over the globe as a whole. The incoming solar radiation, which at the top of the atmosphere in average is 341W/m^2 is set to 100 units. Some of the short-wave (visible) radiation from the sun is reflected from clouds and the surface of the earth. The reflection properties of the surface may vary and is called albedo. The white snow and desert sand has a high albedo, while black soil has a low albedo. Note that only a small fraction of solar radiation is absorbed in the atmosphere. The sun is warming up the ground, which in turn warms the atmosphere. If we did not have the atmosphere, the surface of the earth would have an average temperature of -19°C , in contrast to the present average temperature of about $+13^\circ\text{C}$. This is because of the ability of the atmosphere to absorb long wave radiation (infrared) from the surface of the earth. This is called the greenhouse effect. (Do you know two important greenhouse gases?) The variation of the solar radiation versus time and latitude is illustrated in Fig. 1.2 a and b.

Fig.1.3 shows how the radiation imbalance is distributed versus latitude. As a consequence, heat must be transported pole-wards by the atmosphere and the ocean. How the atmospheric circulation is organised is illustrated in Fig. 1.4.

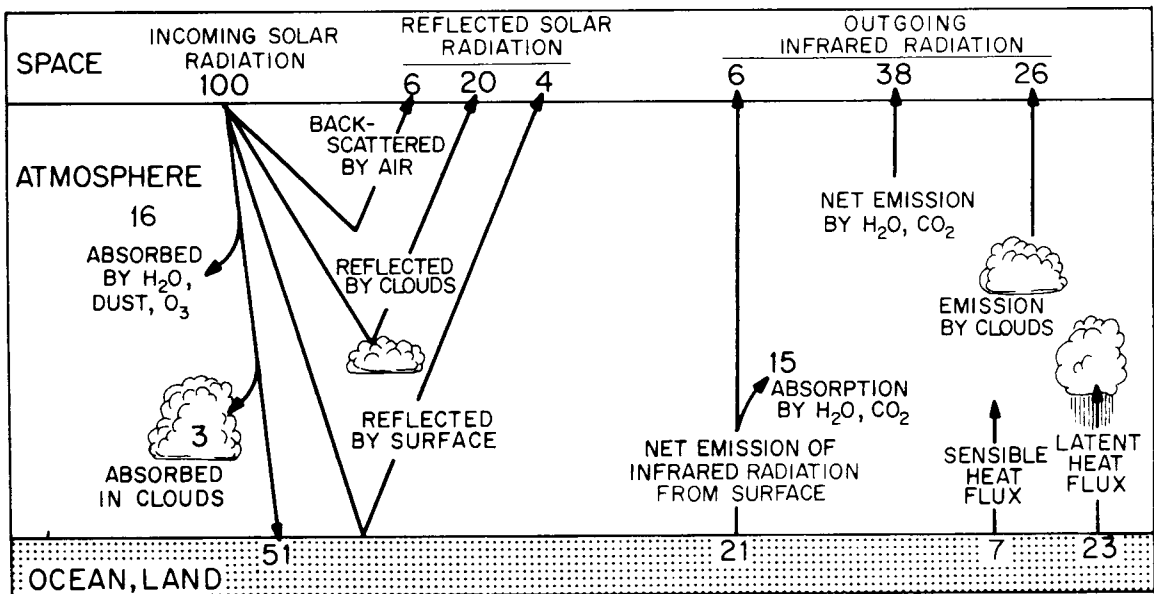


Fig.1.1 Radiation balance for the atmospheric atmosphere. $100=340\text{ W/m}^2$ (From Gill, 1982)

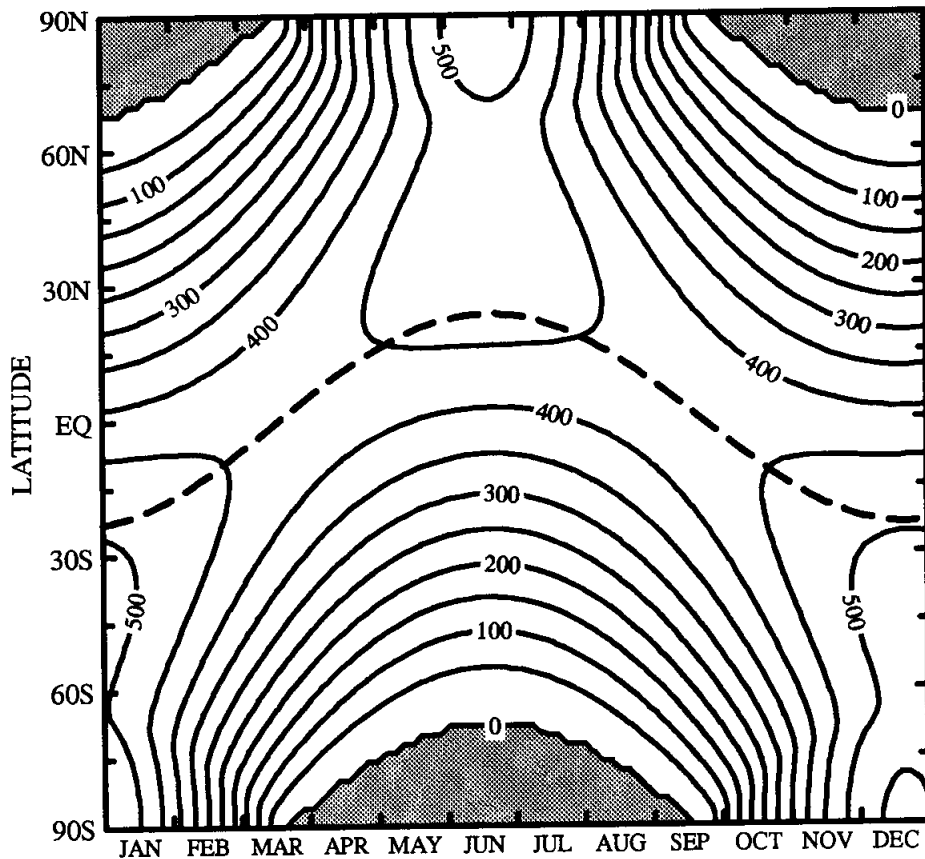


Fig.1.2 a Contour graph of the daily average insolation at the top of the atmosphere as a function of season and latitude. The contour interval is $50\ W\ m^{-2}$. The heavy dashed line indicates the latitude of the sub-solar point at noon.

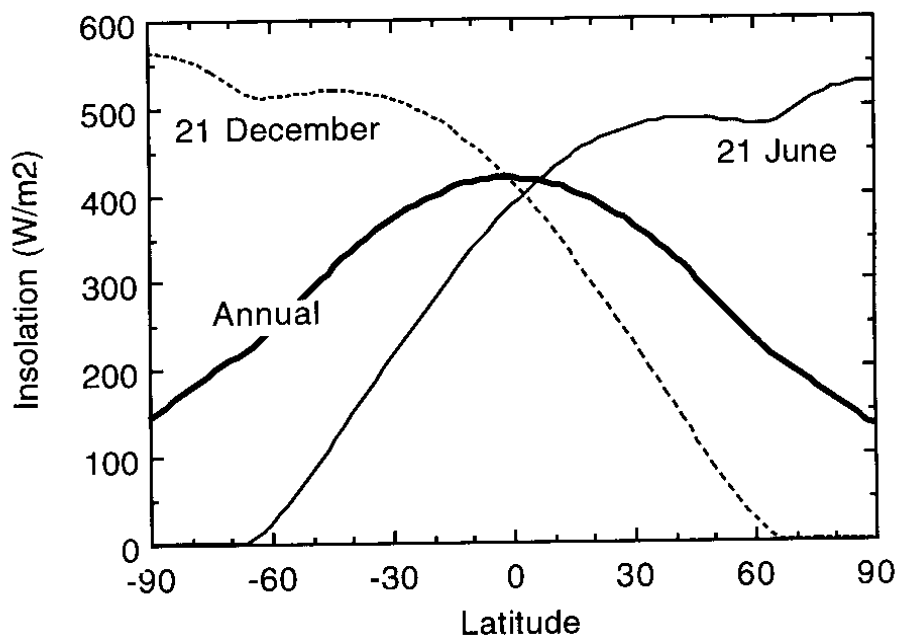


Fig.1.2 b Annual mean and solstice insolation as a function of latitude

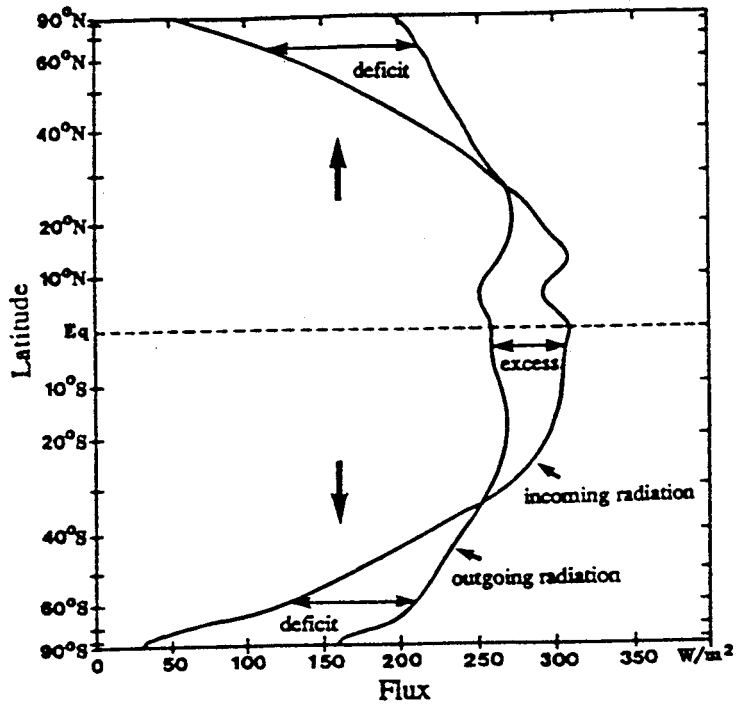


Fig.1.3 Averaged radiation flux by latitude, calculated from satellite data over the period 1974 – 1978. The latitude scale simulates the amount of surface area between latitude bands. Incoming radiation is the shortwave solar radiation absorbed by the earth and atmosphere. Outgoing radiation is the longwave radiation leaving the atmosphere.

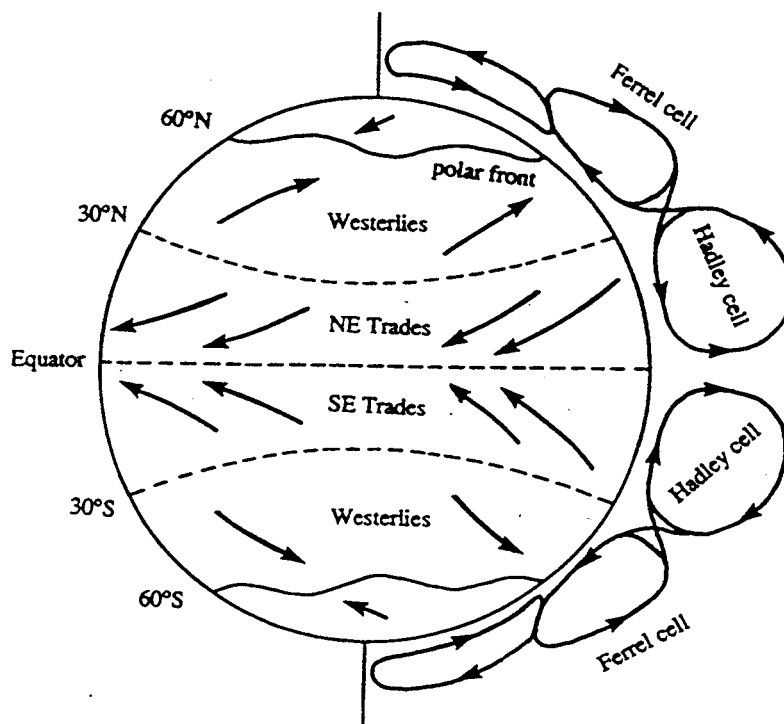


Fig.1.4 Sketch of the general atmospheric circulation, composed of direct (Hadley) and indirect (Ferrel) cells in the meridional direction and of alternating winds in the zonal direction.

2. STATICS

(Ocean at rest)

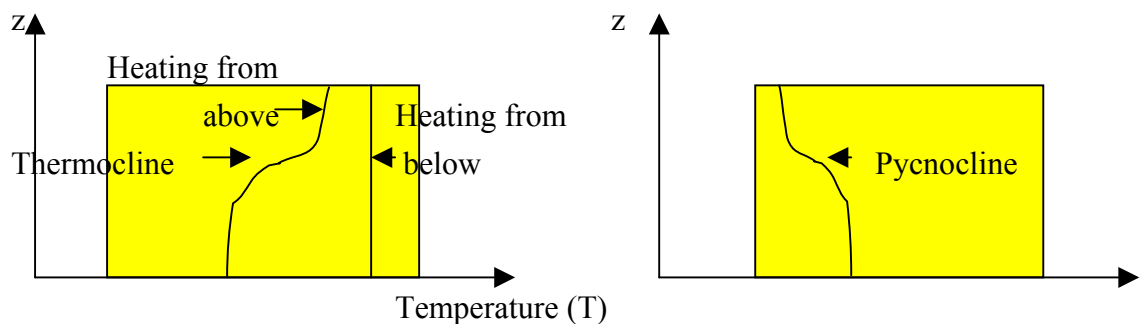
We consider 1 litre of water from the tank. The mass is about 1 kg.

Density (ρ) is defined as mass per volume:

$$\rho = \text{mass/volume} \quad [\text{kg/m}^3]$$

We heat the water in our 1 litre tank. If we heat the water the volume of the water will increase, while the mass is conserved. Thus the density is decreasing.

It makes a difference if we heat the tank from above, or from below. Heating it from below will induce a convection and therefore vertical mixing, so the temperature will not vary with depth. But if we heat from above, the top layer will be less dense, and therefore it may stay put at the surface. In this case we will have a temperature stratification (variation with depth). The region where the temperature stratification is maximum is called the thermocline.



density (ρ)

Fig. 2.1 Temperature in a tank of water heated from below or from above. The corresponding density profile if heated from below is seen on the right hand side. The maximum downward increase on density is called the pycnocline. What does the density stratification look like in the case where it is heated from below?

We may also cool the water in our tank. It is possible to create a temperature (and therefore density) stratification by cooling as well. How?

As you know the water in the ocean is saline. So if we add some salt in our tank, and let it dissolve, the density of the water will increase. A salinity and density

stratification is obtained if we put the salt into the bottom of the tank and let it dissolve there.

If we dissolve 35g salt into the 1 kg of fresh water in the tank the concentration of salt will be 35g/kg or $35\text{g}/1000\text{g} = 35/1000 = 35$ per mille $= 35 \cdot 10^{-3}$

Usually we say that the salinity in this case is 35 psu (Practical Salinity Units)

What is the density of the ocean water?

Of course this will vary with T and S, a few values are given in Table 2.I.

For convenience a quantity called σ_t is introduced:

$$\sigma_t = \rho - 1000 \text{ [kg/m}^3\text{]}$$

Density at atmospheric pressure for some temperatures and salinities:

T	S	ρ	σ_t
0	33	1026.49	26.49
20	33	1023.33	23.23
0	35	1028.10	28.10
20	35	1024.73	24.73

Table I

Stability.

If σ_t increases with depth, we say that the stratification is stable.

σ_t decreases with depth: unstable

$\sigma_t = \text{constant}$ with depth: neutral or indifferent

In the stable case a vertically displaced particle is forced back towards its initial position (Archimedes' principle). This explains how internal waves may occur in the ocean.

As a measure of stability we use $\Delta\sigma_t/\Delta z$

Atmospheric processes which may influence the density:

Evaporation/precipitation

cooling/heating

Exercise 2.1 T – S diagram.

- Table II shows temperature and salinity observations from a hydrographical station. Plot the vertical profiles of temperature and salinity.
- Fig. 2.2 shows a diagram where the isopycnal σ_t is marked. Based on the values shown on the Figure you may also draw the lines for $\sigma_t = 25, 26, 27$ and 28 .
- Plot the pair of T,S values from each depth in Table II on the T-S diagram and draw a line to connect the points.
- Use the T-S diagram to roughly estimate the density at the different depths, and plot the density profile.
- Discuss the stability of the water column, at which depth is the stratification most stable? Least stable?
- We will now see what happens if we mix two water masses of different origin: If a water-mass m_1 with temperature T_1 and salinity S_1 mixes with a water mass m_2 with temperature T_2 and salinity S_2 the temperature T and the salinity S of the mixture is given by:

$$T = (m_1 T_1 + m_2 T_2) / (m_1 + m_2)$$

$$S = (m_1 S_1 + m_2 S_2) / (m_1 + m_2)$$

A water mass with temperature 2°C and salinity 35 mixes with another water mass with temperature 8.7°C and salinity 36. Calculate the T and S values of the mixtures if they mix in the ratio 1:2, 1:1 and 2:1. Plot the “parent” water masses and the values for the different mixtures on the T-S diagram. Compare the density of the mixtures compared to the original water masses.

Table II

Depth meters	Temperature $^\circ\text{C}$	Salinity psu
0	12	33.7
10	12	33.7
20	12	33.8
50	10	33.9
100	8	33.7
200	7	34.1
300	6	34.2
400	5	34.2

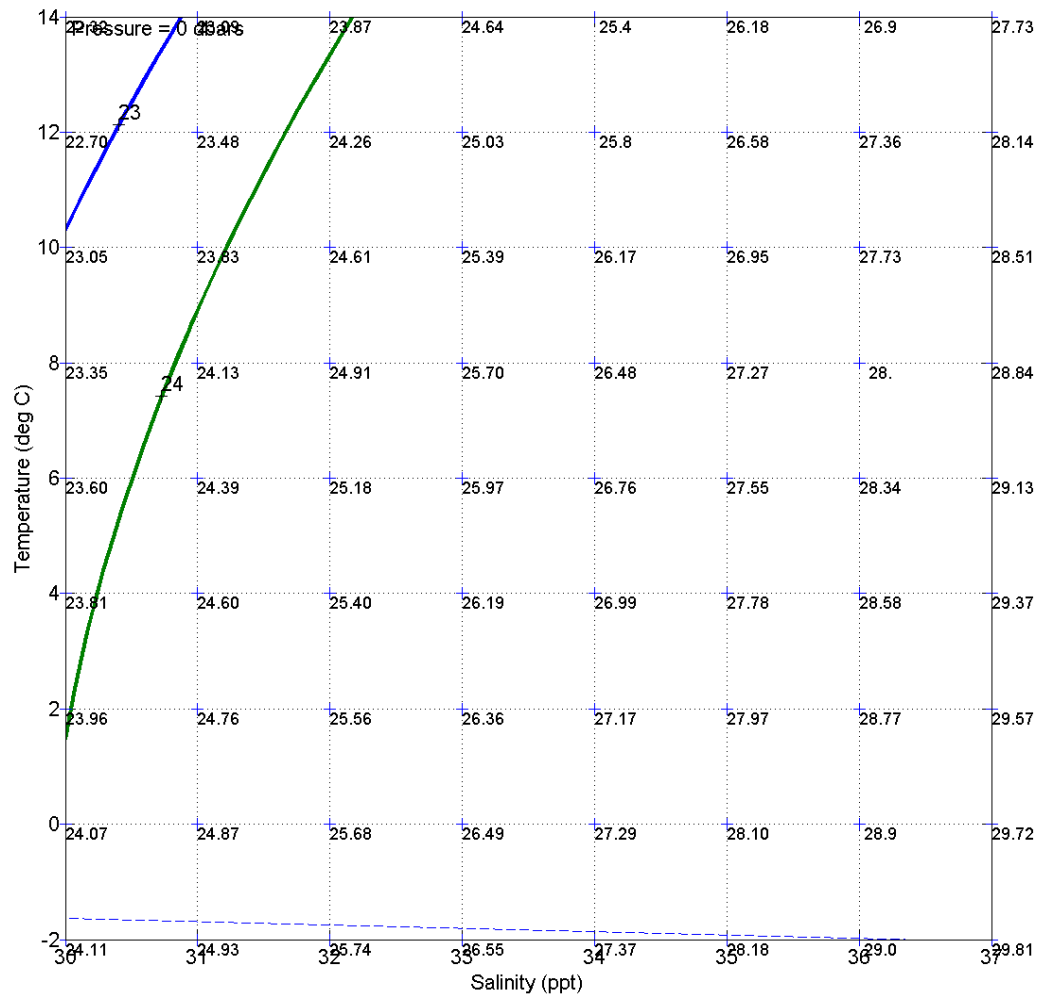


Fig.2.2 T-S diagram.

3. GLOBAL T AND S DISTRIBUTIONS

Considering what we know about the radiation distribution, it is not surprising that the Sea Surface Temperature (SST) is distributed with latitude as demonstrated in Fig. 3.1. Note that SST in the Polar Regions is lower than 0°C , the freezing point of fresh water. This is possible because the freezing point of seawater with a salinity of 34 psu is about -1.9°C .

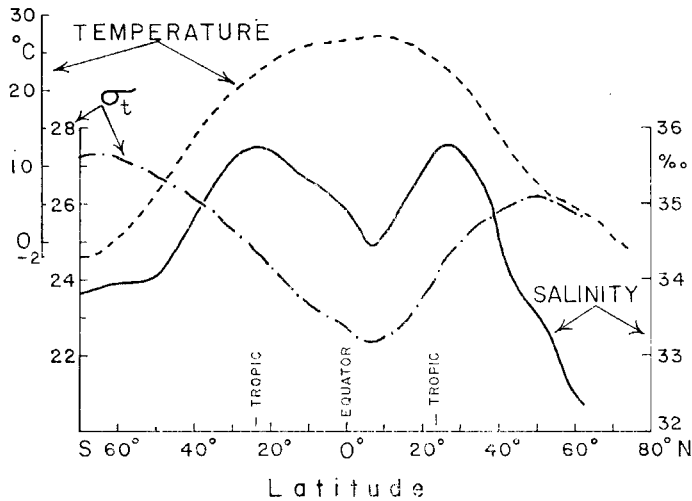


Fig.3.1 Variation with latitude of surface temperature, salinity and density (σ_t) average for all oceans.

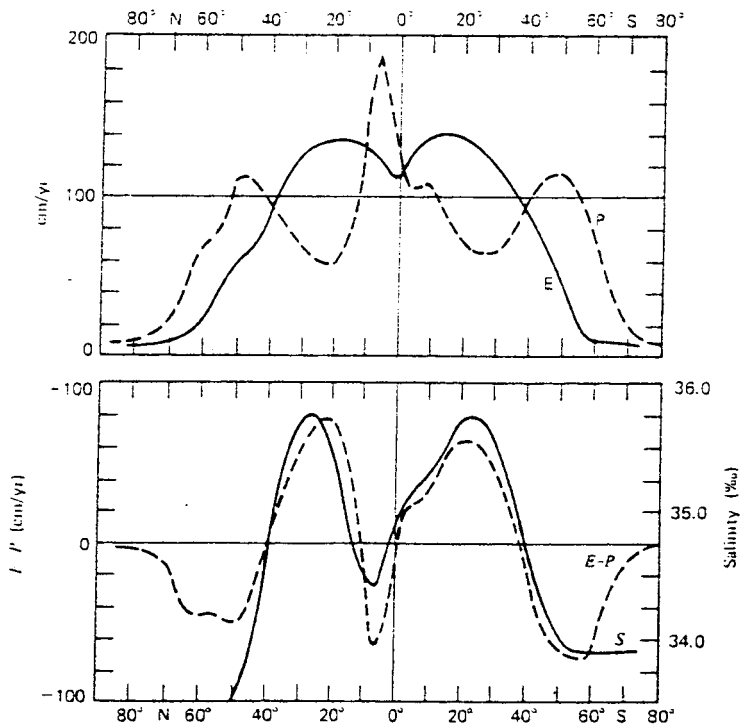


Fig.3.2. Plot of evaporation E and precipitation P as a function of latitude. The difference $(E-P)$ and the surface salinity is given in the lower graph.

The Sea Surface Salinity (SSS) has a more complicated structure, showing a local minimum just north of the Equator. However, this structure in salinity does not change the appearance of the density distribution (σ_t). The SSS structure may be explained by studying the distribution of precipitation and evaporation, see Fig.3.2.

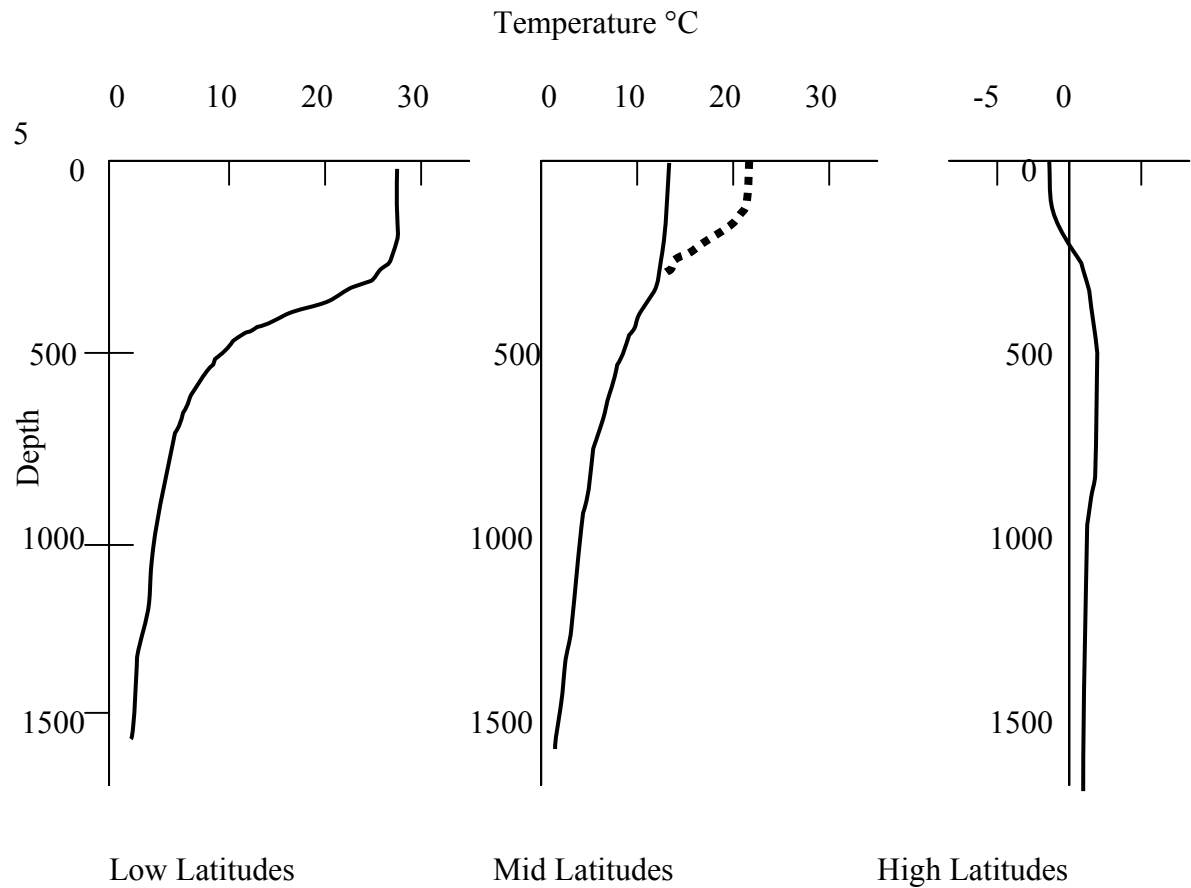
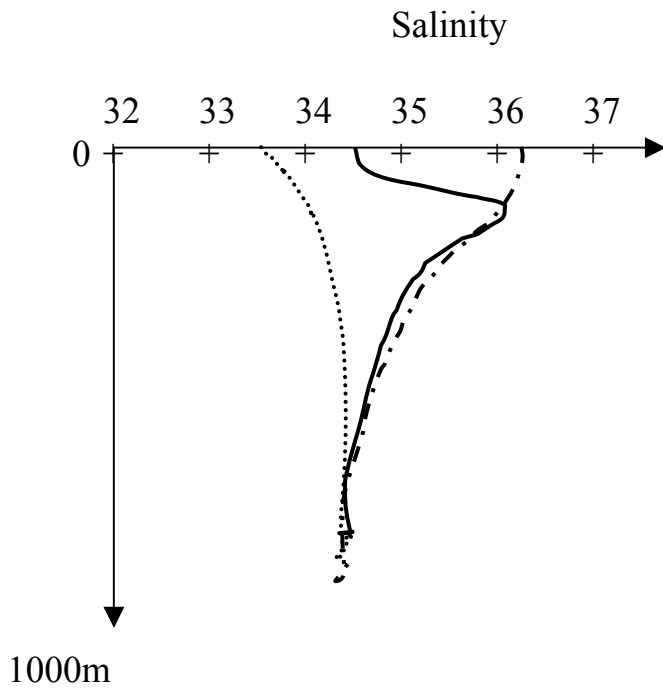


Fig.3.3 Typical vertical temperature profiles in the ocean. Note the deep mixed layer and strong thermocline at low latitudes. The seasonal thermocline is apparent at Mid Latitudes. In Polar Regions T is low at the surface, and does not vary much with depth.

Typical profiles of salinity are shown in Fig. 3.4. Try to identify at which latitudes the different curves belong.



Exercise:
Label the
curves
correctly:

Depth

High ? Latitude
Low? “
Equator ?

Fig. 3.4 Typical mean salinity profiles in the ocean.

Exercise: Which atmospheric processes tend to decrease the stability in the ocean?

In Fig.3.5 typical vertical profiles of density are shown. Note that the stability is very low at high latitudes. This means that deep- and bottom water is formed in Polar Regions. This is in fact a very important driving mechanism of the global circulation; it is called the thermo-haline circulation.

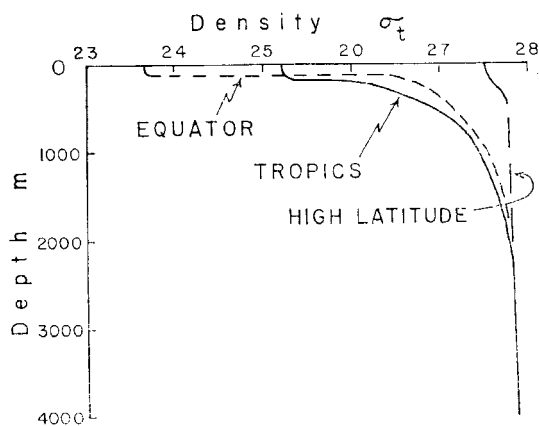


Fig.3.5 Typical density - depth profiles at low and high latitude.

An illustration of the growth and decay of the seasonal thermocline from weather ship station MIKE (66°N, °E) is shown in Fig.3.6. We note how the amplitude of the temperature signal decreases and is delayed with depth.

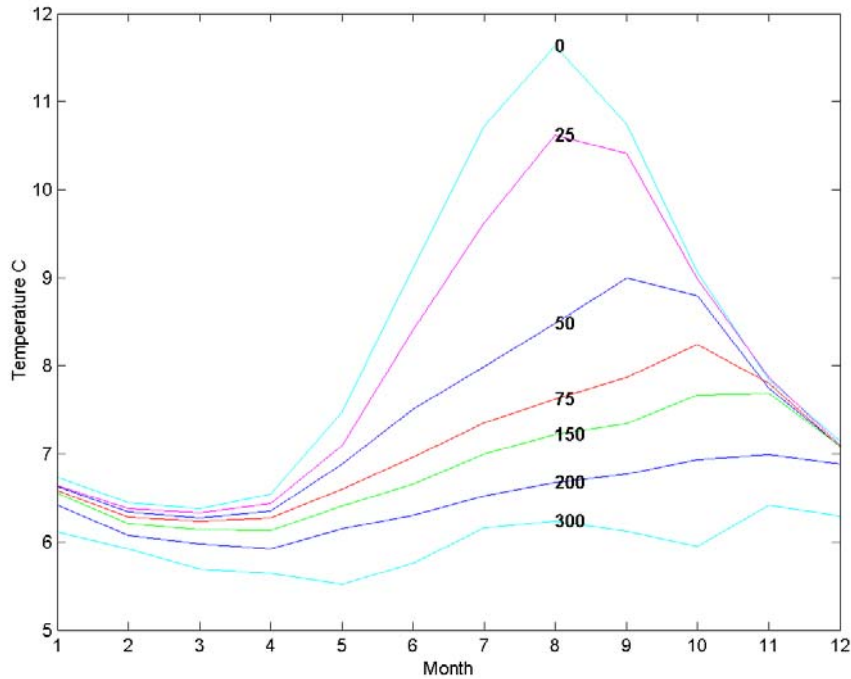


Fig.3.6a. Monthly averages of temperature at different levels at MIKE

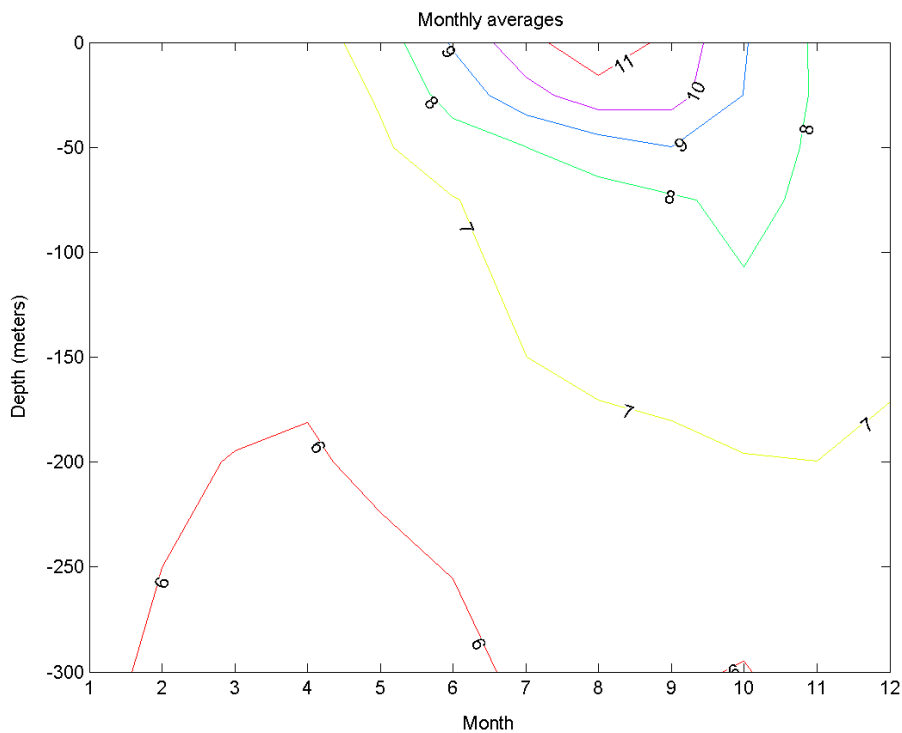


Fig 3.6b. Same as a) but as an isopleth diagram

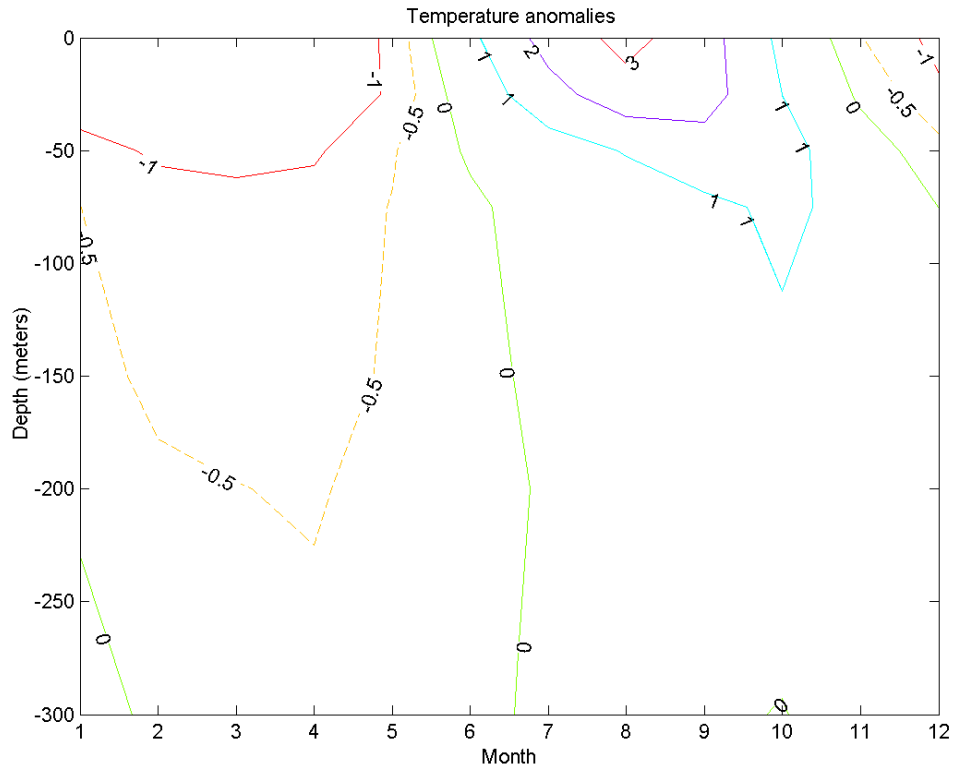


Fig.3.6c Same as b) but the average temperature profile subtracted.

Fig. 3.7 shows the global distribution of SST in August and February. Note the shift in the maximum temperatures with season. Also note the local temperature minimum near Equator in the Eastern Pacific. Namibia seems to be a harsh area, as SST along the coast remains low even in February.

The mean SSS is shown in Fig.3.8. Note the salinity maxima away from Equator. Why is the salinity so low in the Bay of Guinea? Where do you find the absolute maximum SSS?

Vertical N-S sections of T, S and O₂ are shown in Fig.3.9. Note how the dense water formed in the Arctic is flowing southwards as a deep current, characterized by an oxygen maximum. The water formed in the Antarctic is even denser, underlying the water from the north.

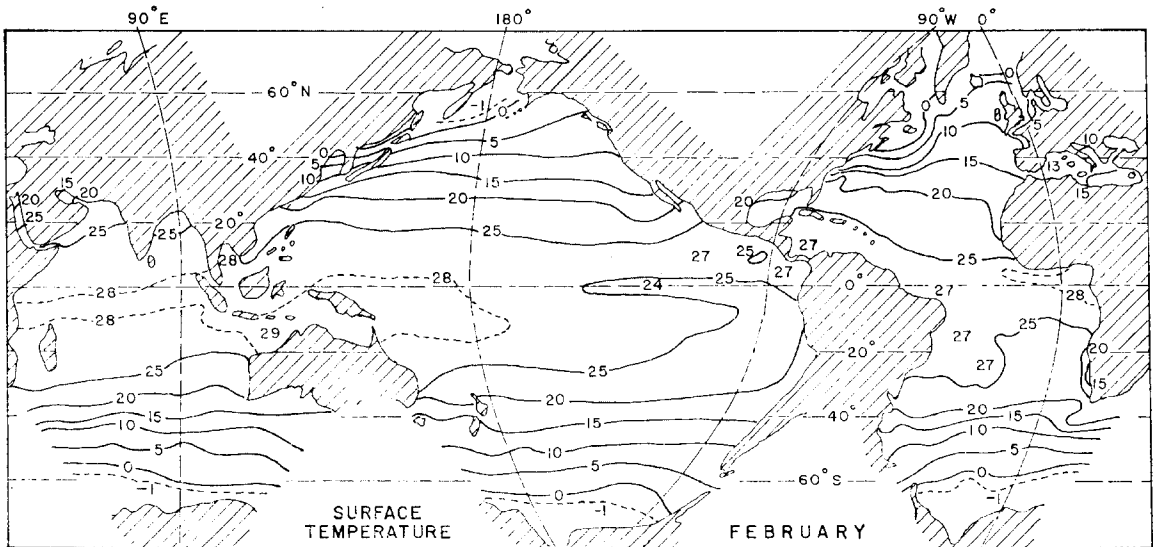
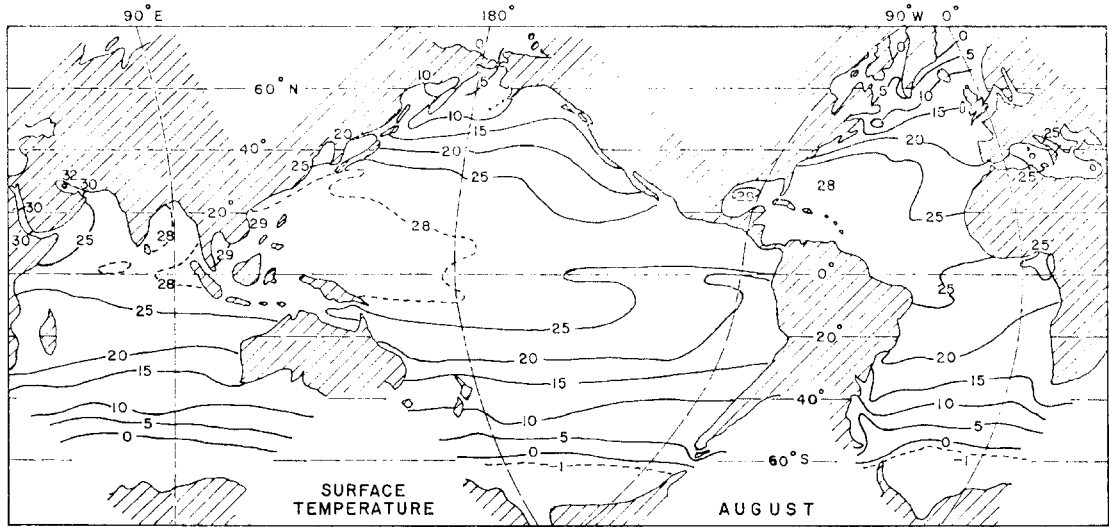


Fig.3.7 Sea Surface Temperature (SST) for August and February

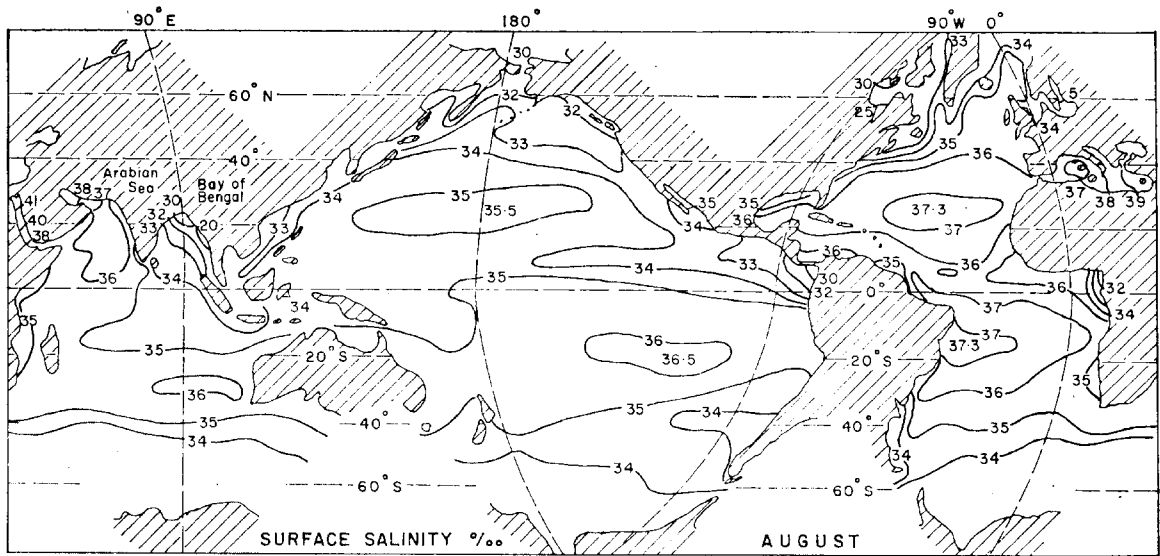


Fig.3.8 Sea Surface Salinity (SSS) of the oceans in August

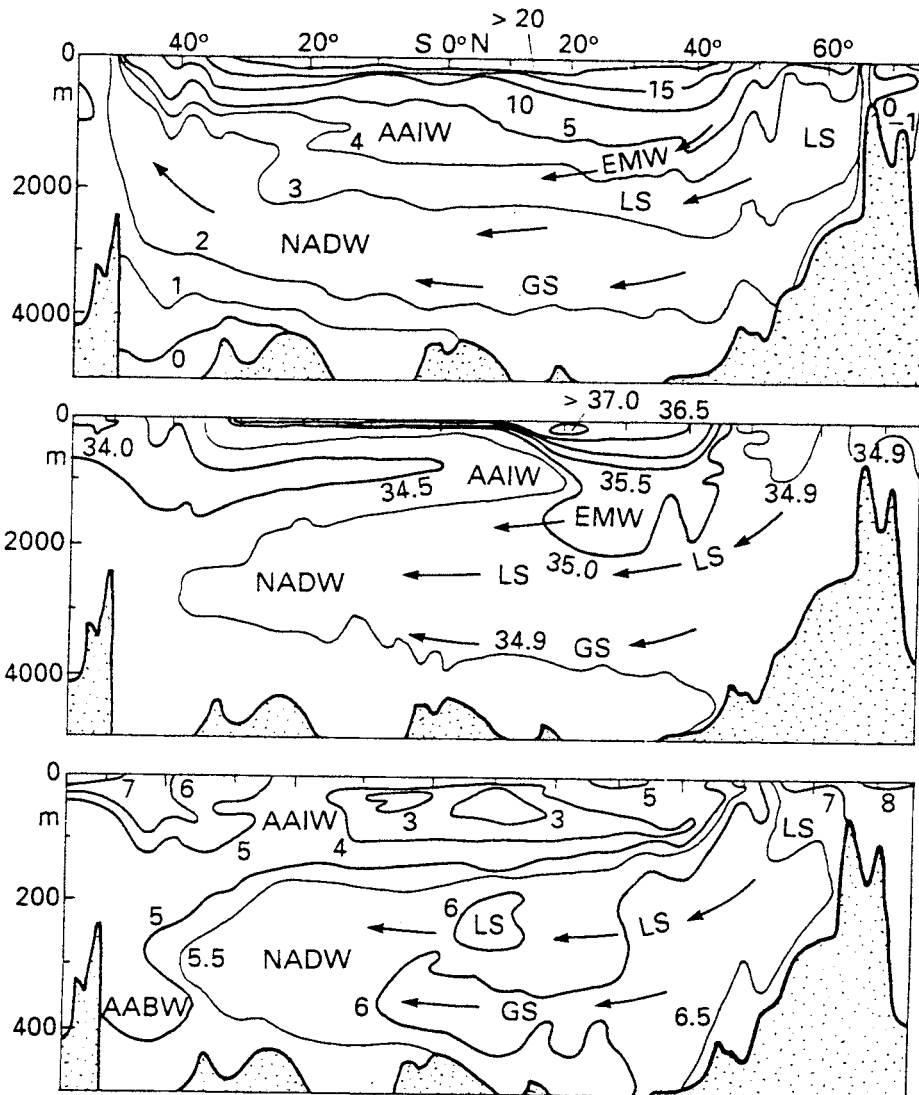


Fig. 3.9 A section through the western basins of the Atlantic Ocean. a) Temperature ($^{\circ}\text{C}$) b) Salinity, c) oxygen (ml/l). AABW: Antarctic Bottom water, AAIW: Antarctic Intermediate Water, NADW: North Atlantic Deep Water originating from the Labrador Sea (LS) or the Greenland Sea (GS) or containing a contribution of Eurafrian Mediterranean Water (EMW).

Light, oxygen and nutrient profiles

We will finish this chapter by having a short look at the absorption of light, and the vertical distribution of oxygen and nutrients in the ocean.

About 50% of the incoming visible light is absorbed in the upper meter of the ocean, see Fig.3.10

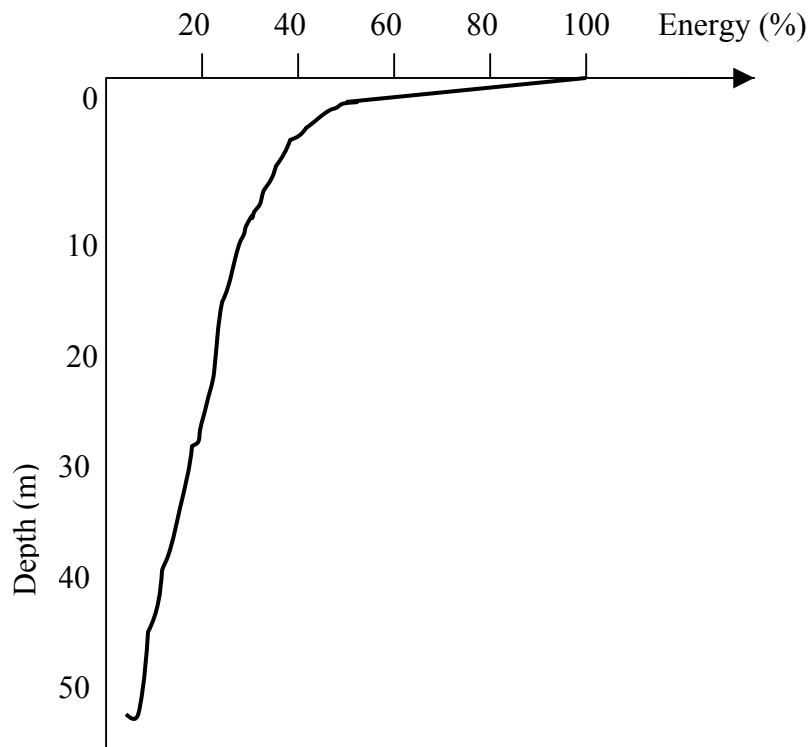


Fig.3.10 Percentage of solar radiation energy reaching different depths (pure water)

Typical profiles of oxygen and nutrients are shown in Fig. 3.11. Note the nutrient depletion in the surface layer, and the O₂ minimum below.

TV

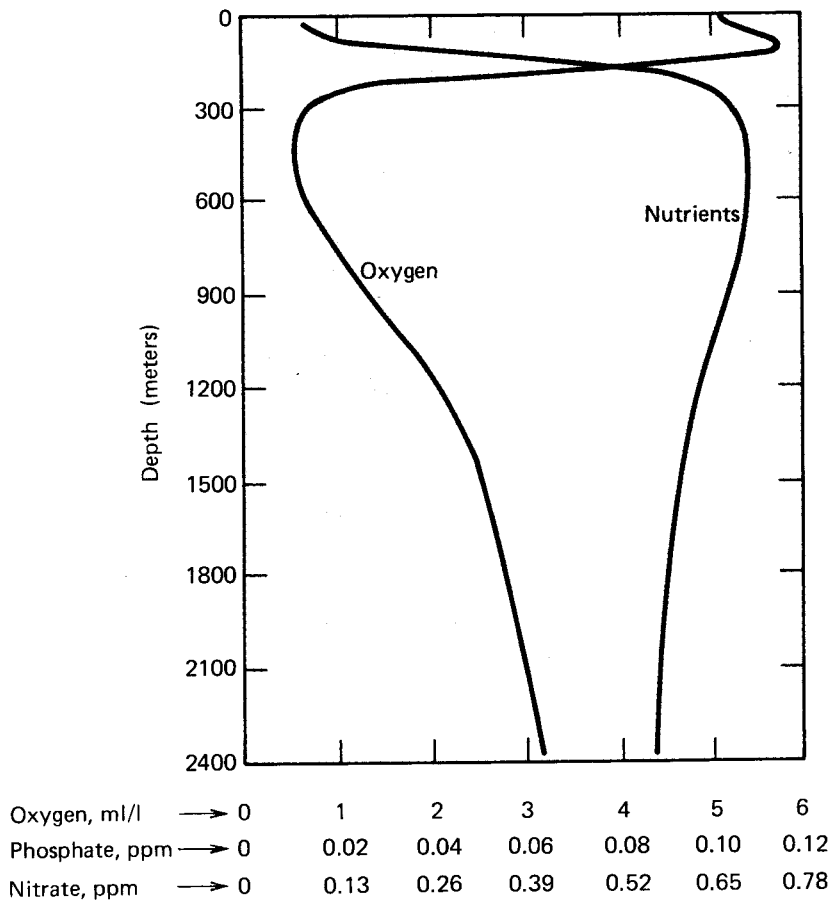


Fig.3.11. Plots of nutrient and oxygen concentration as a function of depth for the North Pacific off Southern California. The effect of the plants depleting the nutrients and enriching the oxygen in the surface waters is clearly displayed. Also, the oxygen minimum region testifies to the animal activity there.

4. DYNAMICS

(Ocean currents)

Horizontal and vertical currents

Let us as consider a model of the ocean as a rectangular basin (Fig.4.1). The typical width of the ocean $L=1000\text{km}$, and a typical depth $D=1\text{km}$.

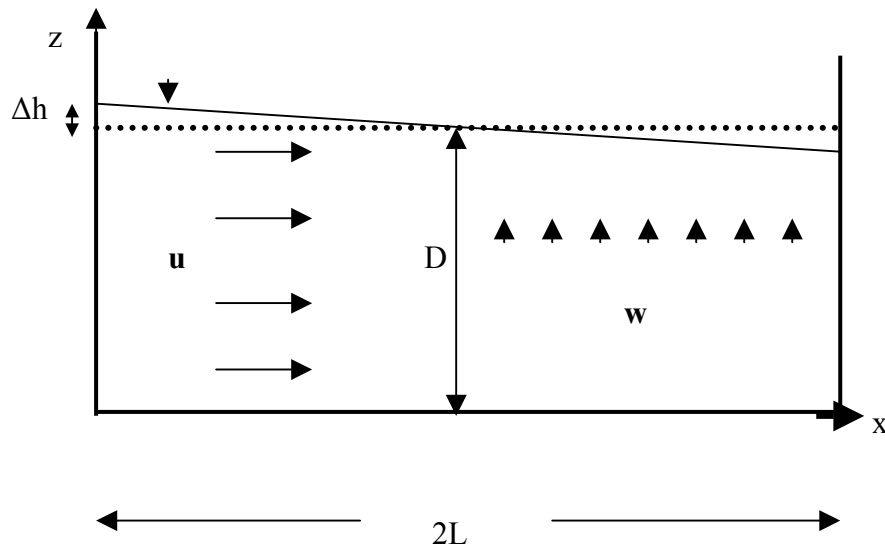


Fig.4.1. Model of an ocean basin

If we measure currents in the ocean, we find that a typical velocity would be $u=10\text{cm/s}$, but stronger currents up to 1m/s is occasionally found in the upper layer, while deep currents seldom exceed 1 cm/s .

What is the typical vertical velocity?

From continuity considerations (Fig.4.1) we note that

$$u \cdot D \approx w \cdot L$$

so

$$w \approx u \cdot D / L = u \cdot 10^{-3} = 10^{-2} \text{ cm/s}$$

Therefore: Ocean currents are mainly horizontal. Vertical velocities are weak, but still very important for biological production. (Why?)

Convergence and divergence

Vertical velocities are usually associated with convergence or divergence, see Fig.4.2

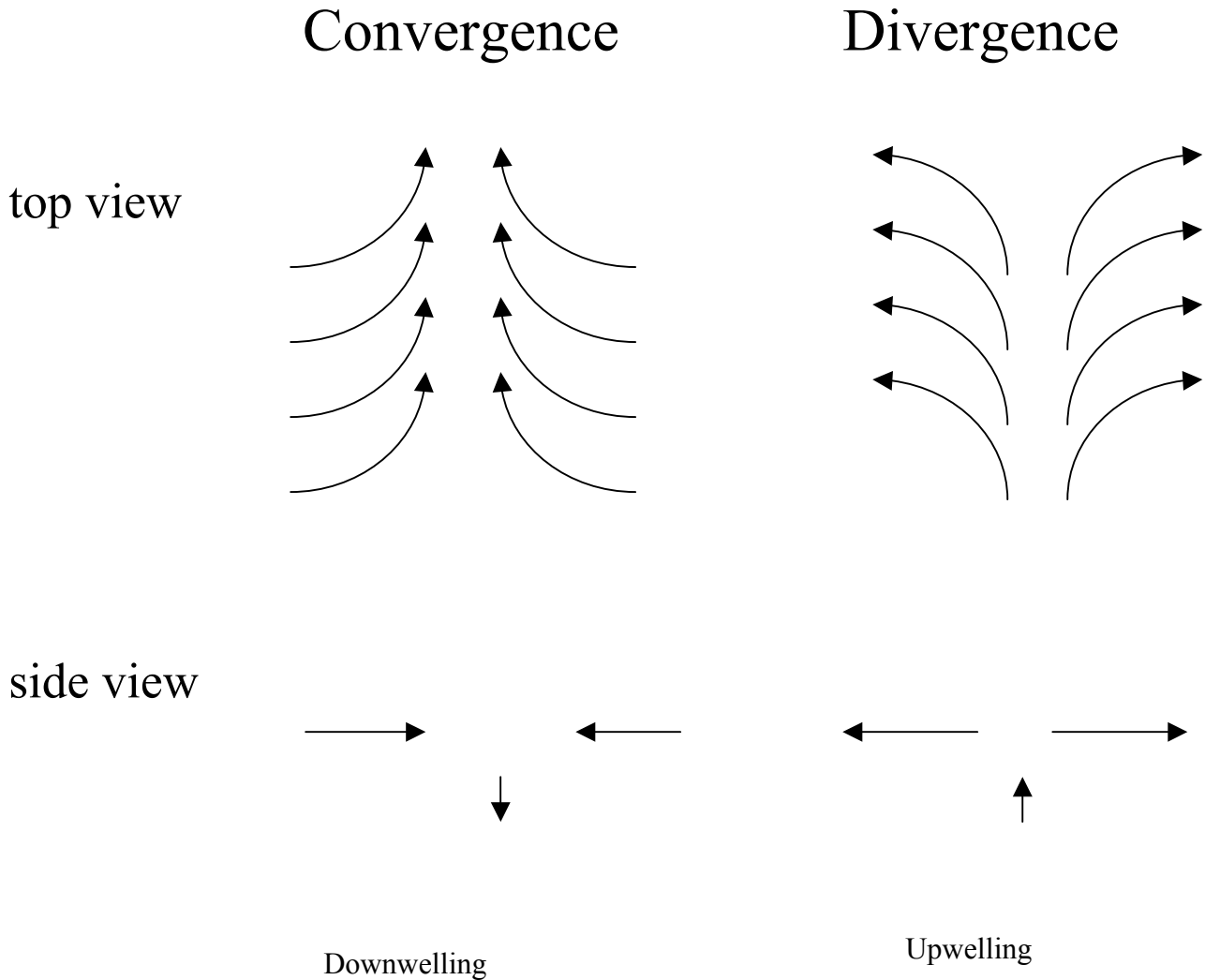


Fig. 4.2. Downwelling is associated with convergence, and upwelling by divergence

Forces driving the ocean currents

Currents in the ocean obey Newtons 1st and 2nd law of motion:

Newtons 2nd law:

The sum of forces (\vec{K}) acting on a body with mass m gives an acceleration (\vec{a}) according to the formula:

$$\vec{K} = m\vec{a}$$

Newtons 1st law: If the sum of forces is zero;

$$\vec{K} = 0$$

Then the motion will continue at a constant speed and direction ($\vec{v} = \text{const.}$)

For the large scale ocean currents the acceleration is usually small, so the sum of forces add up to zero ($\vec{K} = 0$).

Important forces are:

Pressure force: \vec{P}

Friction: \vec{F} Friction is due to turbulence. Friction only occurs when the water is moving. No motion; no friction. (Secondary force)

Coriolis \vec{C} Due to the rotation of the earth. No motion no Coriolis force. (Secondary force)

Gravity $\vec{g} = g\vec{k}$ Only vertical component. $g=9.81\text{m/s}^2$

So in the ocean we usually have

$$\vec{P} + \vec{F} + \vec{C} = 0$$

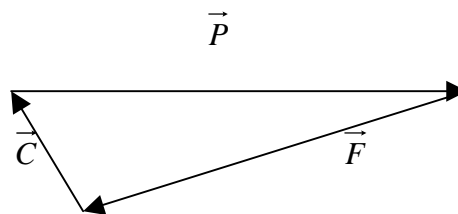


Fig. 4.3. Illustrating the balance between 3 forces.

As the currents are mainly horizontal, these are horizontal forces.

Hydrostatic pressure and pressure force

Pressure mean force per unit area. It may be calculated as follows: Consider a column of water with density ρ , height h and area A . See Fig.4.4

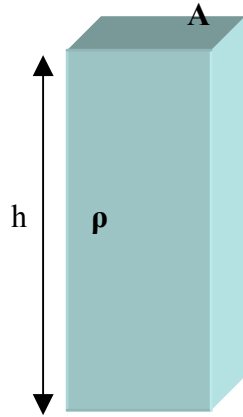


Fig. 4.4 Considering a column of water

The weight of the water column is mass x gravity: mg

The mass is $m = \rho hA$

so the weight becomes ρhAg

So this is the force due to the weight of the water at depth h

As pressure (p) is the force per unit area, it may be calculated as:

$$p = \rho hg$$

This is called the hydrostatic equation and is used to calculate the hydrostatic pressure.

Pressure forces arise when the pressure varies horizontally. Let us go back to the ocean basin, Fig.4.1, and consider the pressure at the horizontal bottom when the surface is tilting:

The pressure at the bottom in the centre ($x=L$) is:

$$p(x=L) = \rho gD$$

while the pressure at western coast $p(x=0)$ is

$$p(x=0) = \rho g(D + \Delta h)$$

Thus the difference in the pressure over the distance L is

$$\Delta p = p(x=0) - p(x=L) = \rho g(D + \Delta h) - \rho gD = \rho g\Delta h$$

and the pressure gradient is (defining $L = \Delta x$):

$$\Delta p / \Delta x = (p(x=L) - p(x=0)) / L = \rho g\Delta h / \Delta x.$$

The water is pushed from high pressure to low pressure and the pressure force is proportional (but opposite direction) to the horizontal pressure gradient. So for this example the pressure force will yield an acceleration given by:

$$\vec{P} = -\frac{1}{\rho} \frac{\Delta p}{\Delta x} \hat{i} = -g \frac{\Delta h}{\Delta x} \hat{i}$$

where \hat{i} is the unit vector along the x-axes.

Note that for this model, when we have a constant density, the same pressure force will be obtained at any level. Try for instance to calculate the pressure gradient at the level $z=D/2$ (half way down, and at $z=0$ (the surface)). This means that the pressure force is constant through the depth, and is only due to the tilt of the surface. This will then give the same acceleration at all depths, as indicated on Fig. 4.1.

We will now consider an ocean where the surface is horizontal, but the density is varying horizontally, see Fig. 4.5.

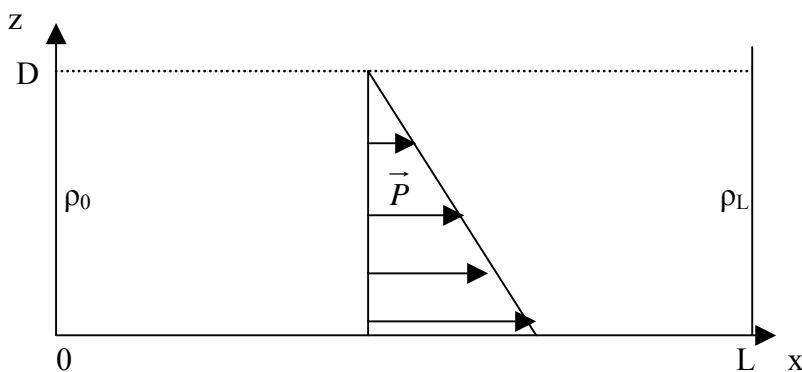


Fig. 4.5. An ocean model where the density varies horizontally (with the distance x).

Let us assume that the density at the western coast ($x = 0$) is higher than at the eastern coast ($x = L$);

$$\rho_0 > \rho_L$$

We now calculate the pressure at the bottom at $x=0$ and at $x=L$

$$p(x=0) = p_0 = \rho_0 g D$$

and

$$p(x=L) = p_L = \rho_L g D$$

The pressure difference becomes

$$p_0 - p_L = \Delta p = (\rho_0 - \rho_L) g D$$

In this case the pressure force becomes:

$$\vec{P} = -\frac{1}{\rho} \frac{\Delta p}{\Delta x} \hat{i} = -\frac{\rho_0 - \rho_L}{\rho} \frac{g D}{\Delta x} \hat{i} = -\frac{\Delta \rho}{\Delta x} \frac{g D}{\rho} \hat{i}$$

So the pressure force is proportional (but opposite direction) to the horizontal density gradient.

Note that in this case the depth D comes into play. Which means that at the depth $z=D/2$ we will only have half the pressure force, and at the surface there is no pressure force, see Fig.4.5

Coriolis force

According to Newton's first law an object continues at constant speed if there is no forces acting on it. Imagine you are in the centre of a merry-go-round and you roll a ball outwards. Disregarding friction the ball will follow a straight line, but for you who are sitting in the centre and rotating with the merry-go-round it will appear like the ball is deflected, see Fig.4.6. This apparent force deflecting it is called the Coriolis force. If there is no motion, there is no Coriolis force, (secondary force). The Coriolis force acts perpendicular to the velocity.

If we consider a plane on the rotating earth we notice that on the Poles a plane makes one full rotation in 24 hours, but at the Equator it does not rotate at all. Let Ω be the rotation rate of the planet:

$$\Omega = 360^\circ/24\text{hours} = 2\pi/24\text{h}$$

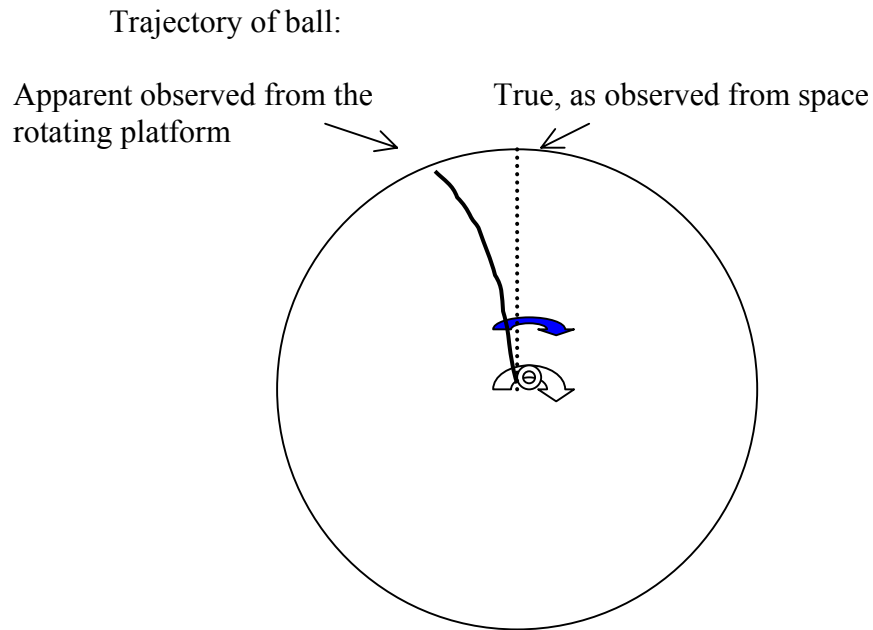


Fig.4.6. The Coriolis force demonstrated at a rotating plane.

The rotation rate of the plane on the planet will vary with latitude (ϕ) and the local rotation rate is given as $\Omega \sin(\phi)$. The Coriolis force is proportional to two times the local rotation rate, which is called the Coriolis parameter f :

$$f = 2\Omega \sin(\phi).$$

Thus the magnitude of the Coriolis force proportional to the speed (v), and depending of the sign of $\sin(\phi)$ it is directed to the right of the velocity at the northern hemisphere, and to the left at the southern hemisphere(see Fig.4.7):

$$C = fv$$

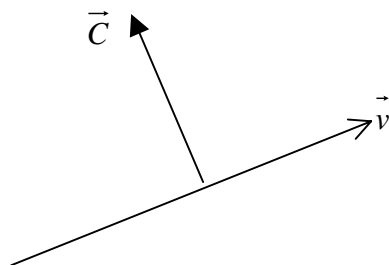


Fig. 4.7. Illustrating the Coriolis force at the Southern hemisphere

The Coriolis term is important for motions with large time- (a few days ore more) and space- (10 – 50km or more) scales.

Geostrophic currents

Measurements show that for large-scale, deep ocean currents only the pressure force and the Coriolis force are important. So the large scale ocean current are described by the geostrophic balance:

$$\vec{P} + \vec{C} = 0$$

and the resulting velocity given by $v=C/f$ is called the geostrophic velocity.

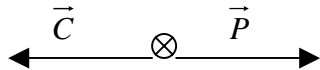


Fig.4.8. The geostrophic balance

The Coriolis force as illustrated in Fig.4.8 is created by a velocity into the paper plane (southern hemisphere).

Going back to our model ocean basin let us consider the ocean with constant density and a tilting surface:

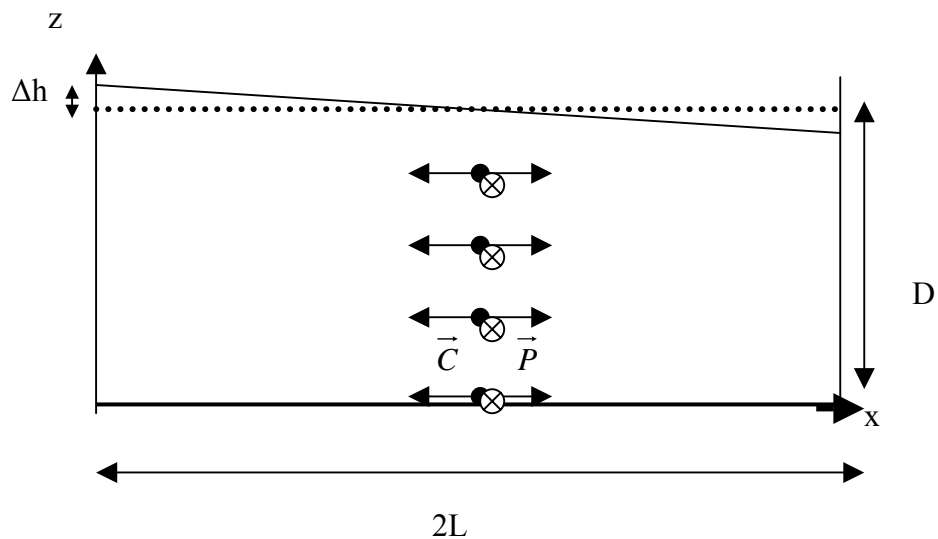


Fig.4.9. Geostrophic model in a constant density ocean.

In an ocean with constant density and a tilting surface, the pressure force, and therefore the associated Coriolis force according to the Geostrophic model, are constant with depth. If the coordinate x is east, then the geostrophic current in Fig.4.9 is northwards (along positive y -axis): see Fig.4.10.

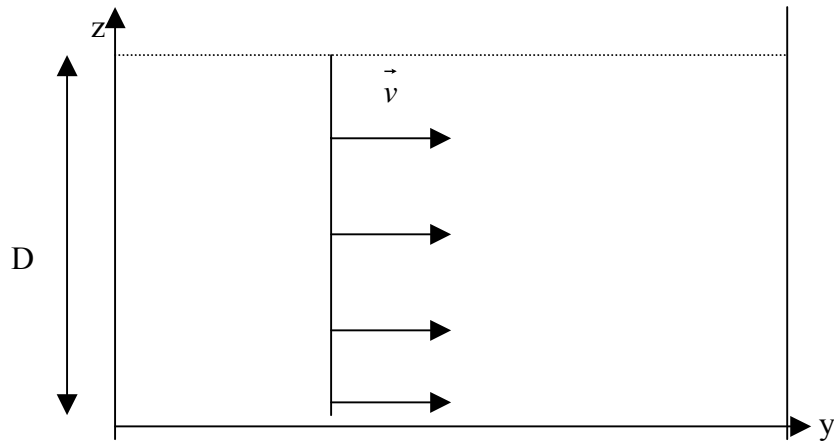


Fig. 4.10. The geostrophic velocity associated with the model in Fig.4.9

The corresponding geostrophic balance and the geostrophic velocity associated with the model with horizontal surface and a horizontal density gradient are illustrated in Fig.4.11 and 4.12.

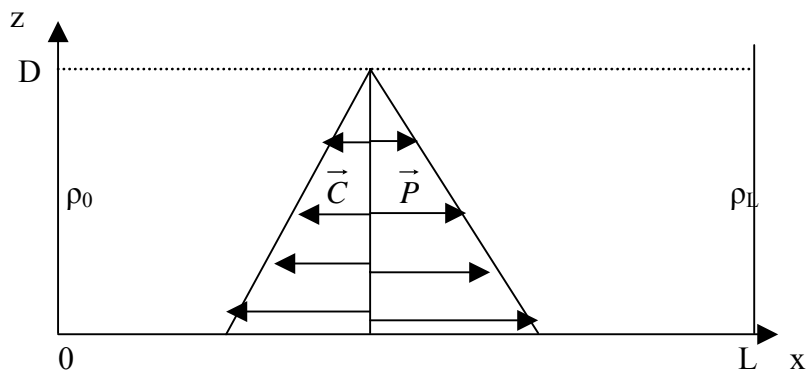


Fig. 4.11. The geostrophic balance in an ocean with a horizontal density gradient.

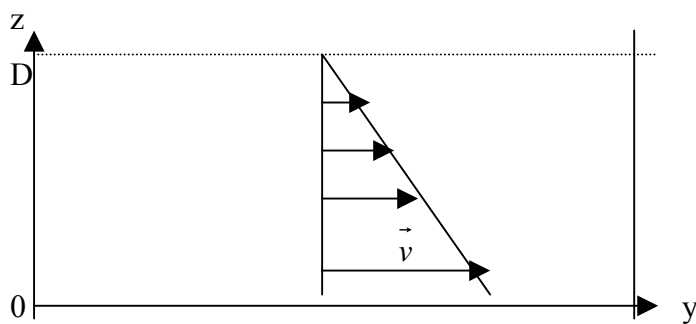
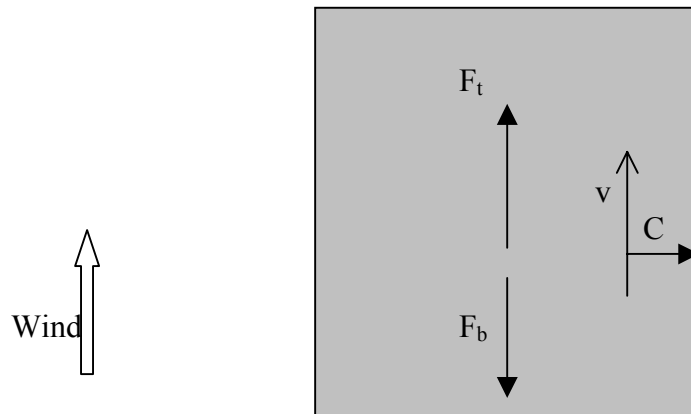


Fig. 4.12. The geostrophic velocity associated with the model in Fig.4.11

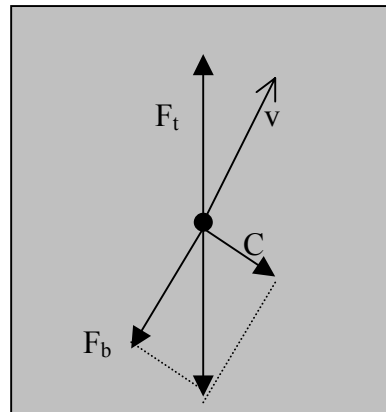
Friction

Wind driven currents.

Friction only occurs when there is motion. The friction is directed opposite to the current velocity. Consider a water layer near the surface of the ocean, or more easy to imagine: an ice floe with wind blowing above (F_t) and friction is felt from the water layer below (F_b), see Fig. 4.13.



a)



b)

Fig. 4.13 Wind driven ice floe. A balance is possible when the ice moves at an angle to the right of the wind (Northern hemisphere)

From Fig.4.13 a) we note there is no force to balance the Coriolis force. A balance is only possible if the ice floe moves at angle to the wind, to the right on the northern hemisphere. So for wind-driven currents the balance of forces is:

$$\vec{F}_t + \vec{F}_b + \vec{C} = 0$$

A theoretical model called the Ekman model shows that the mass transport in the wind-influence layer is directed perpendicular to the wind, to the right (left) on the northern (southern) hemisphere. This is called the Ekman Transport (M_E), see Fig.4.14

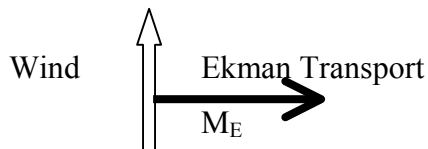


Fig 4.14 The Ekman Transport on the Northern Hemisphere.

This has an important implication: when the wind is blowing parallel to the coast it will give up- or down-welling.

Exercise: Which wind direction would cause up-welling near the Namibian coast? The Mozambican coast? The Norwegian coast?

The ocean surface currents are mainly wind-driven. The currents will be modified by continents, bottom topography, heating/cooling and tides.

In Fig.4.15 a very simple wind driven ocean model is shown. We assume straight north – south oriented coasts, and we simplify the wind patterns, compare Fig.1.4. The results of this model indicate that near Equator where the Coriolis parameter is small and therefore the Coriolis force weak, the currents follow mainly the wind direction. When it hits the western coast, it is split in a northern and a southern branch. But as we now are going away from Equator, the current tends to be bent back eastwards because of the Coriolis force. On its way eastwards, the Coriolis force is still strong, such that a return flow towards the Equator is taking place along the whole track, leading to a much weaker boundary current at the east coast than the west coast. This is called westward intensification. In summary, the solution is big gyres at both sides of Equator and a strong western boundary current.

What does the real world look like? In Fig. 4.16 a more realistic wind pattern and sea level pressure map is given, and in Fig. 4.17 and 4.18 the surface circulation of the Atlantic

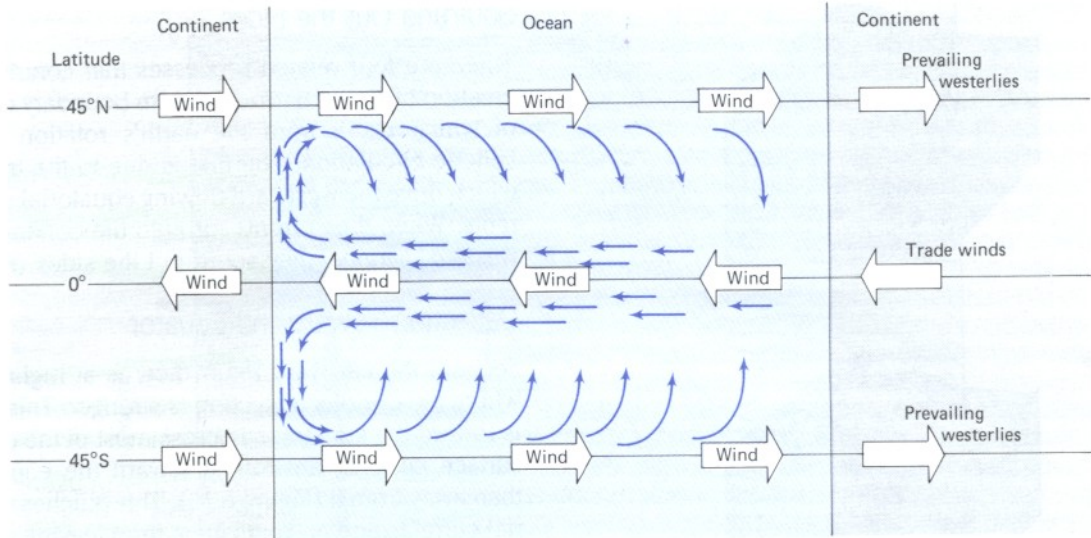


Fig.4.15. Wind and ocean circulation patterns in an idealised ocean basin. Because the Coriolis force is stronger at higher latitudes, the eastward flowing waters are strongly deflected back toward the equator. Consequently, surface waters tend to flow toward the equator everywhere except along the western edge of the ocean.

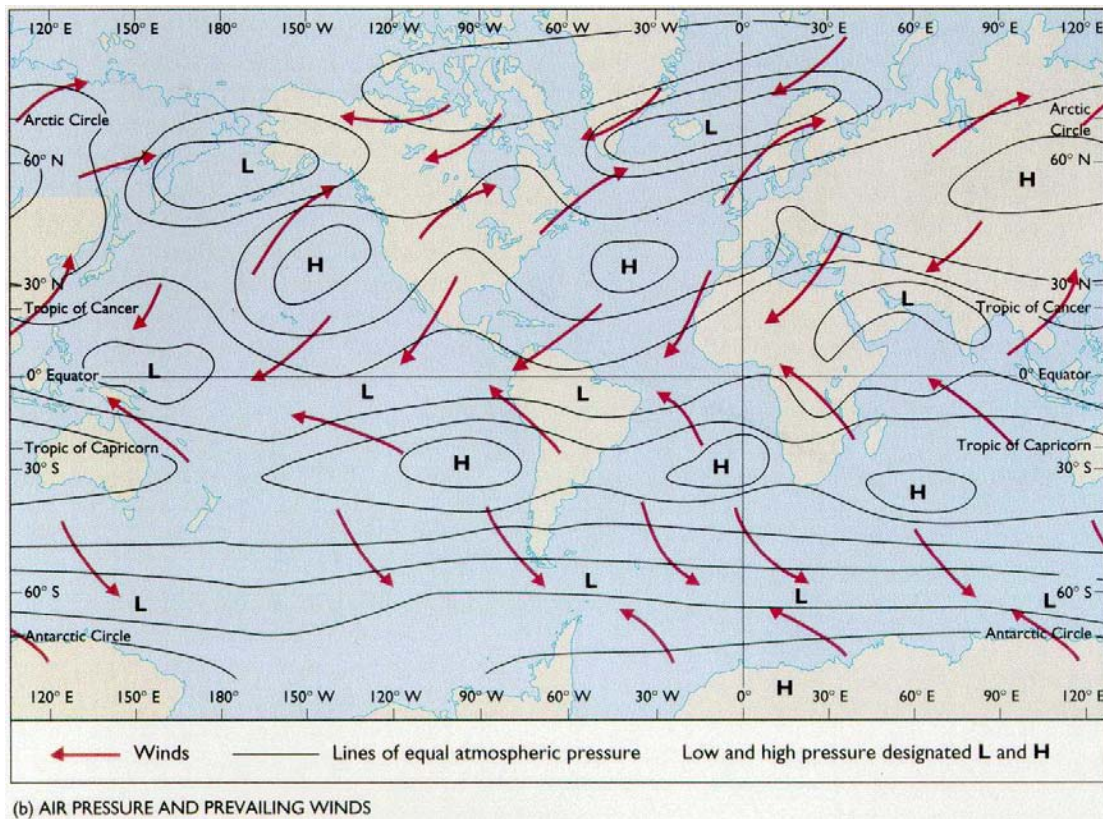


Fig.4.16. Yearly averaged sea level pressures, indicating the regions of high and low pressures and the general pattern of the surface winds.

and the Pacific oceans are given. We recognise both in the Atlantic and in the Pacific some of the main features of our simple model, indicating that the basic physics we assumed is correct.

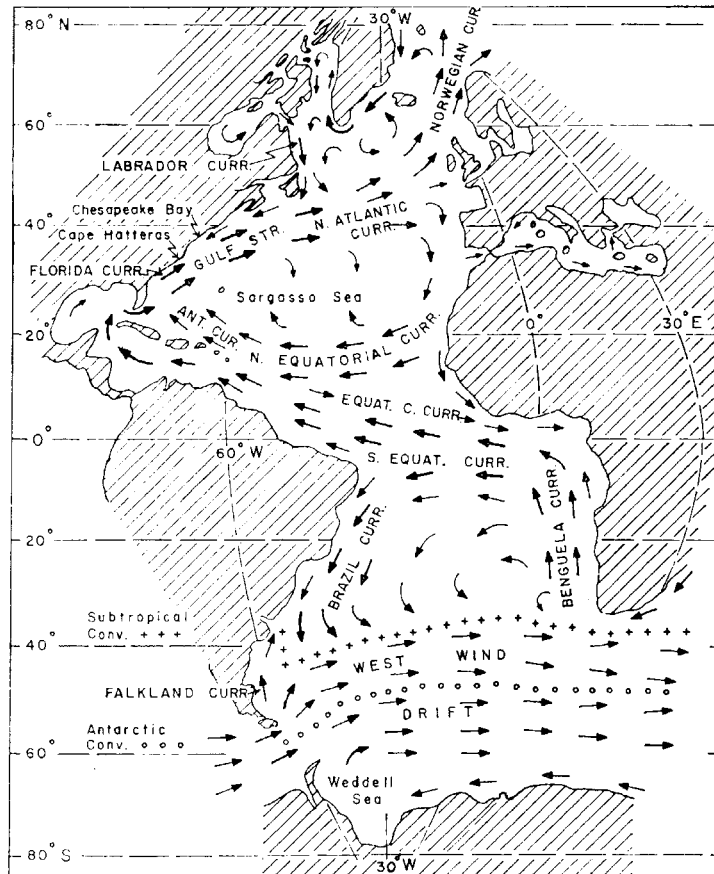


Fig.4.17. Atlantic Ocean – surface circulation

In Fig. 4.19 the surface circulation of the Indian Ocean is shown. It has been necessary to show this on two maps, because the currents reverse with the season. These are known as Monsoon Currents. In November – March (northern hemisphere winter) the earth cools down much faster than the ocean. Thus the air has a tendency to rise above the ocean and leading to winds from the land towards the ocean. Because of the Coriolis force the current tends to go to the right of the wind, leading to a westward current during this season. In the summer the situation is opposite.

Exercise: Mention some reasons why the land warms up faster than the ocean in the summer.

The surface current systems in the Antarctic and the Arctic are shown in Figs. 4.20 and 4.21.

Before we end this chapter let us have a look at the distribution of phosphatic deposits, an indicator of primary production, see Fig.4.22. Compare the surface pressure map and associated winds in Fig.4.16. Note that the Benguela is one of the 4 major upwelling systems.

Exercise: Point out and name the other 3.

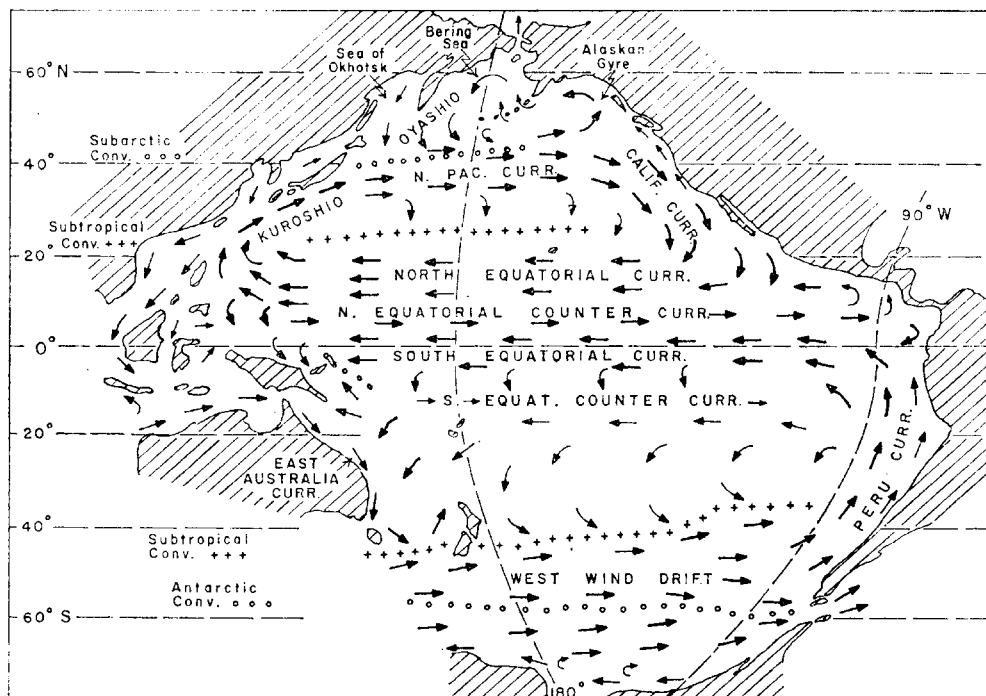


Fig. 4.18. Pacific Ocean – surface circulation.

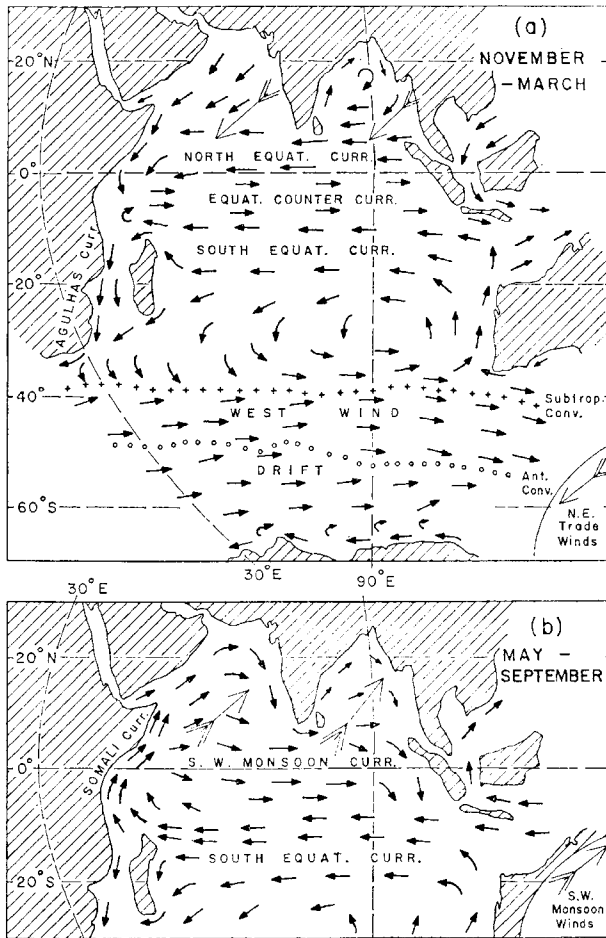


Fig.4.19 Indian Ocean – surface circulation

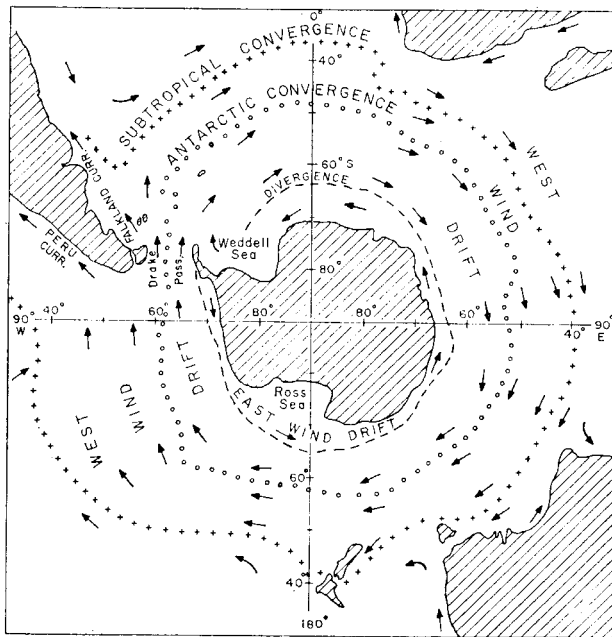


Fig.4.20 Antarctic Ocean – surface circulation

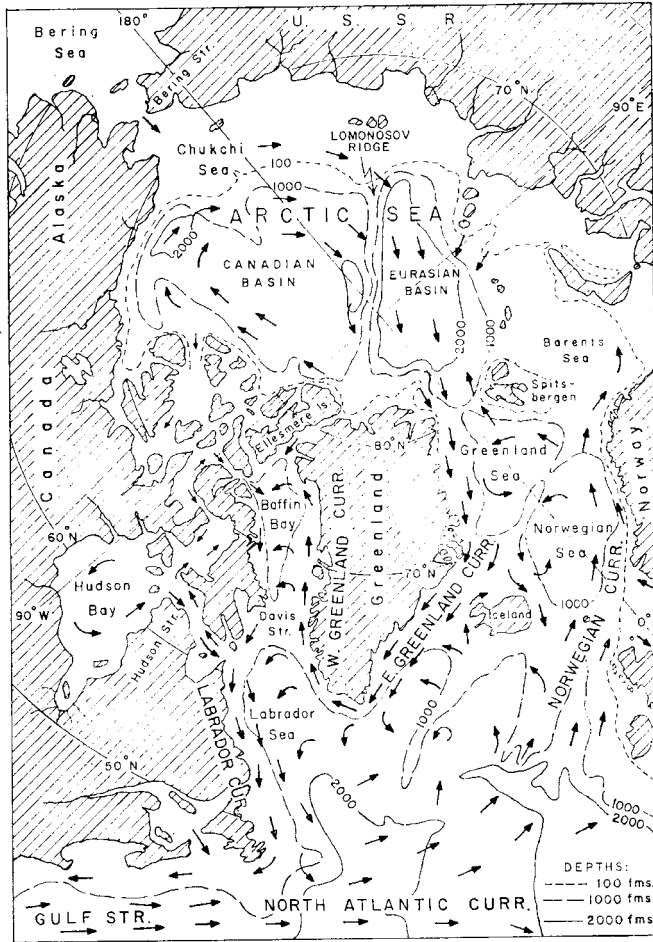


Fig.4.21 Arctic Ocean – surface circulation

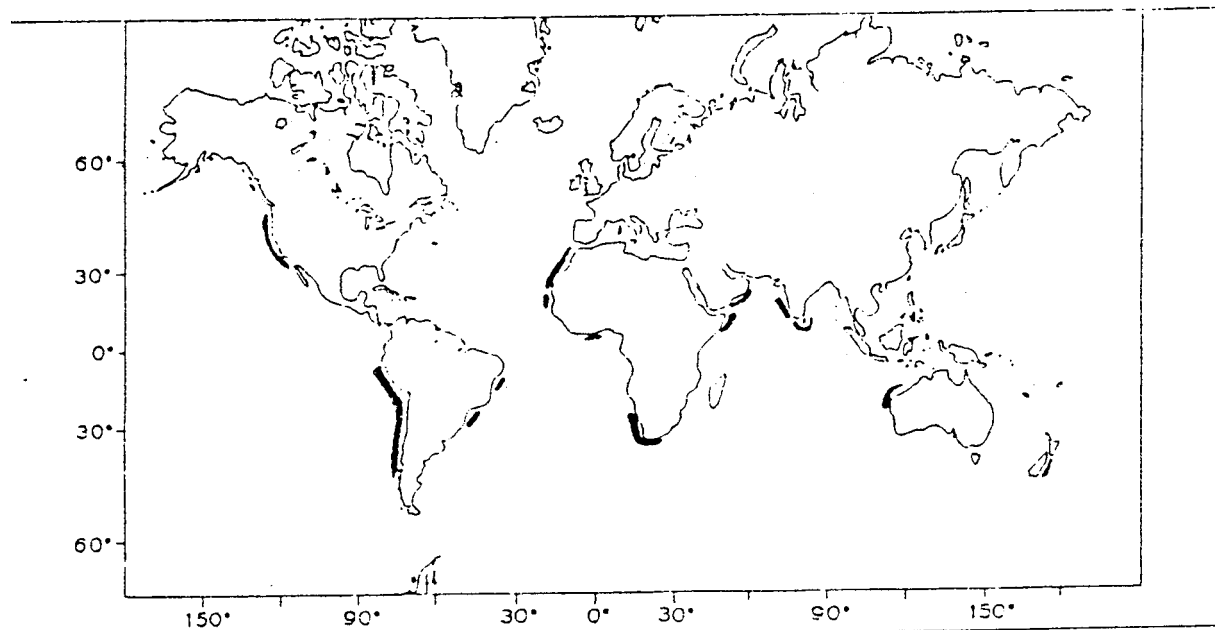


Fig.4.22 Distribution of phosphatic deposits, as an indicator for the major upwelling systems

5.WAVES

Waves are characterized as periodic in space: Fig.5.1,

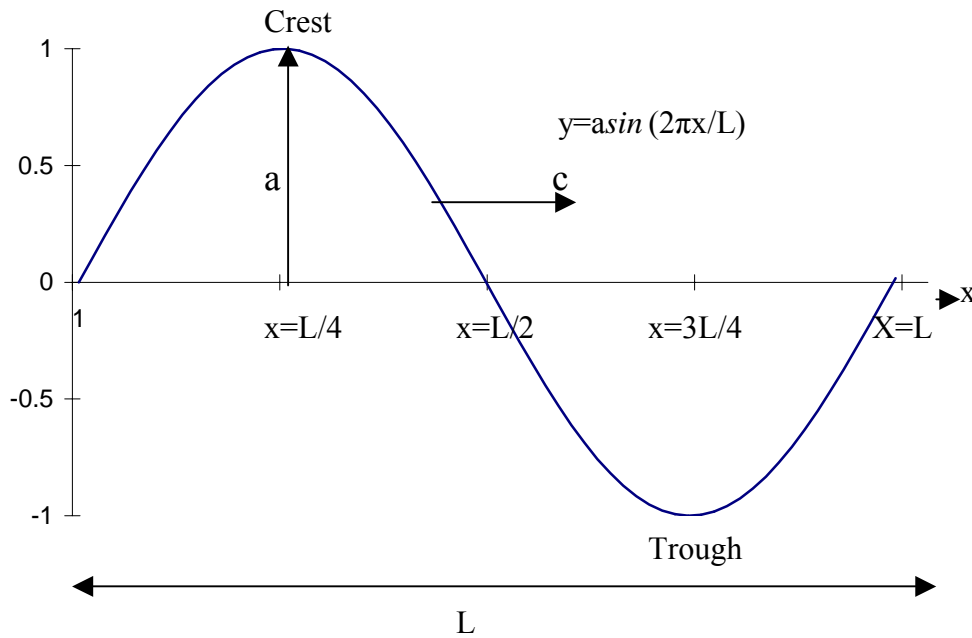


Fig.5.1 Showing wave periodicity in space

and time Fig.5.2

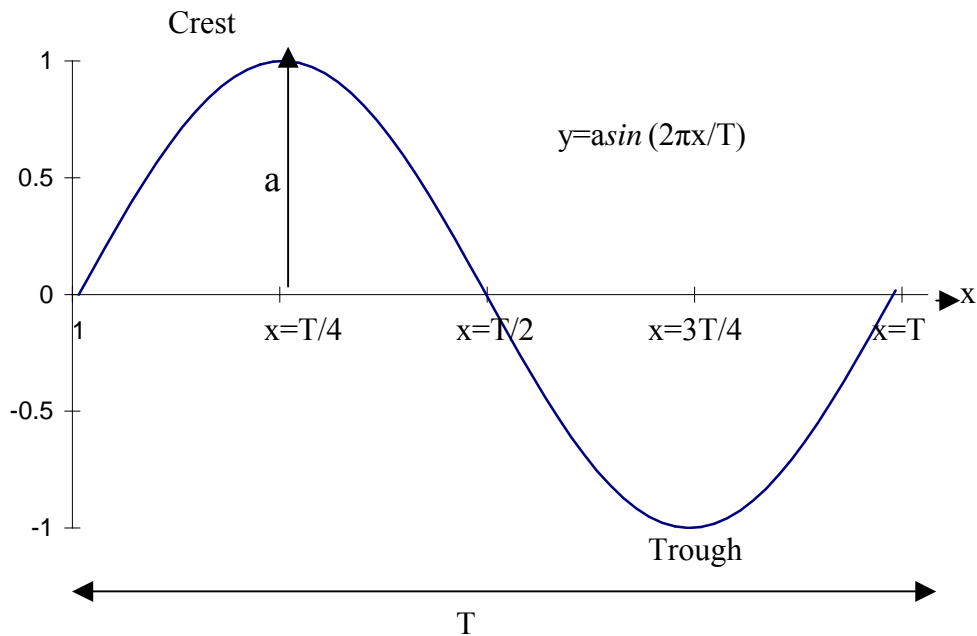


Fig.5.2 Showing wave periodicity in time

Here is a : amplitude, L : wavelength and T : period. Frequency is defined as $2\pi/T$ and the speed of the wave (phase speed c) is given as $c=L/T$.

Superposition

One interesting feature of waves is that the travel of any one wave train is completely independent of the motion of any other. Even when two or more different wave train happens to be passing through the same region at the same time, each continues on its way as if the others were not there. Any portion of the real ocean surface is a collage of many different wave trains of all shapes and sizes and going many different directions. Each of these continues on its way across the ocean, unaffected by the others.

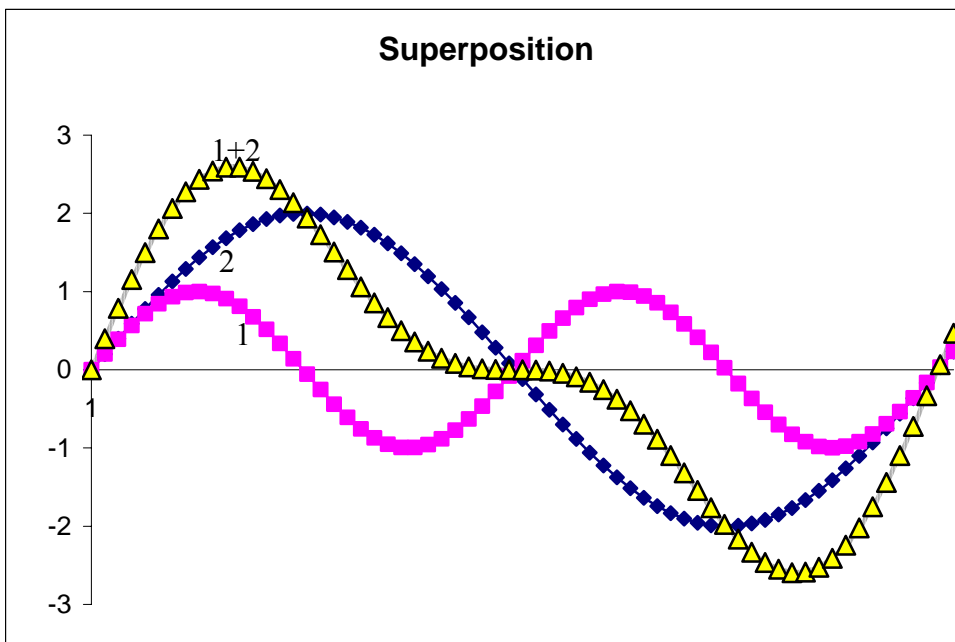
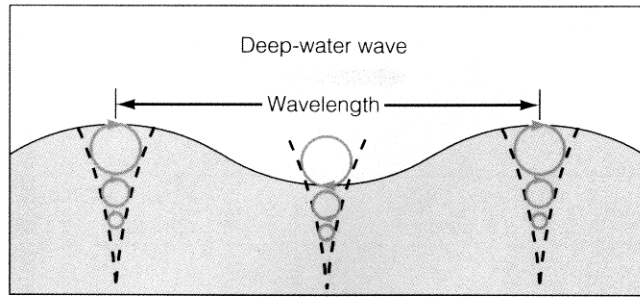


Fig.5.3. In the superposition of waves, the resulting displacement is the sum of the two others.

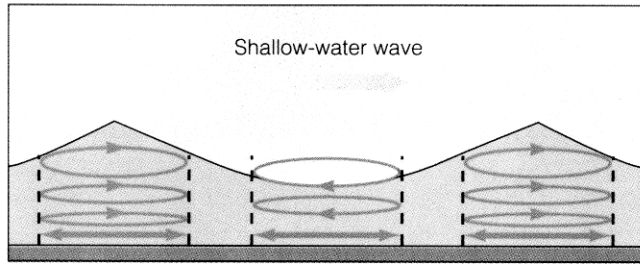
Deep (short) and shallow (long) water waves

Waves are called long or short if the wavelength is long or short compared to the water depth. Short waves have maximum amplitudes near the surface, and the amplitude vanishes a few wavelengths down in the ocean, while long waves have more or less constant amplitude with depth, see Fig.5.4

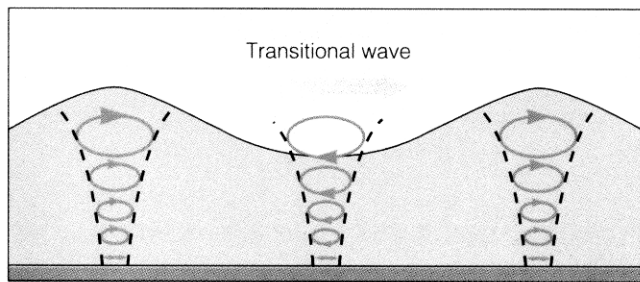
For long waves the phase speed is given by $c = \sqrt{gh}$, where h is the ocean depth. Remember this formula only applies for waves which are long compared to the ocean depth where they occur.



a Depth $\geq \frac{1}{2}$ wavelength



b Depth $\leq \frac{1}{20}$ wavelength



c $\frac{1}{20}$ wavelength \leq depth $\leq \frac{1}{2}$ wavelength

Fig.5.4. Deep and shallow water waves

Energy spectrum

Waves are classified according to their wavelengths and periods. An example of relative energy of various wave types is shown in Fig 5.5. Note that the highest energy is found for wind waves. At low frequencies, the period (and wavelength) is so long that the rotation of the earth have importance (planetary waves). One example of a planetary wave is the equatorial Kelvin wave, which is believed to cause the El Niño phenomenon.

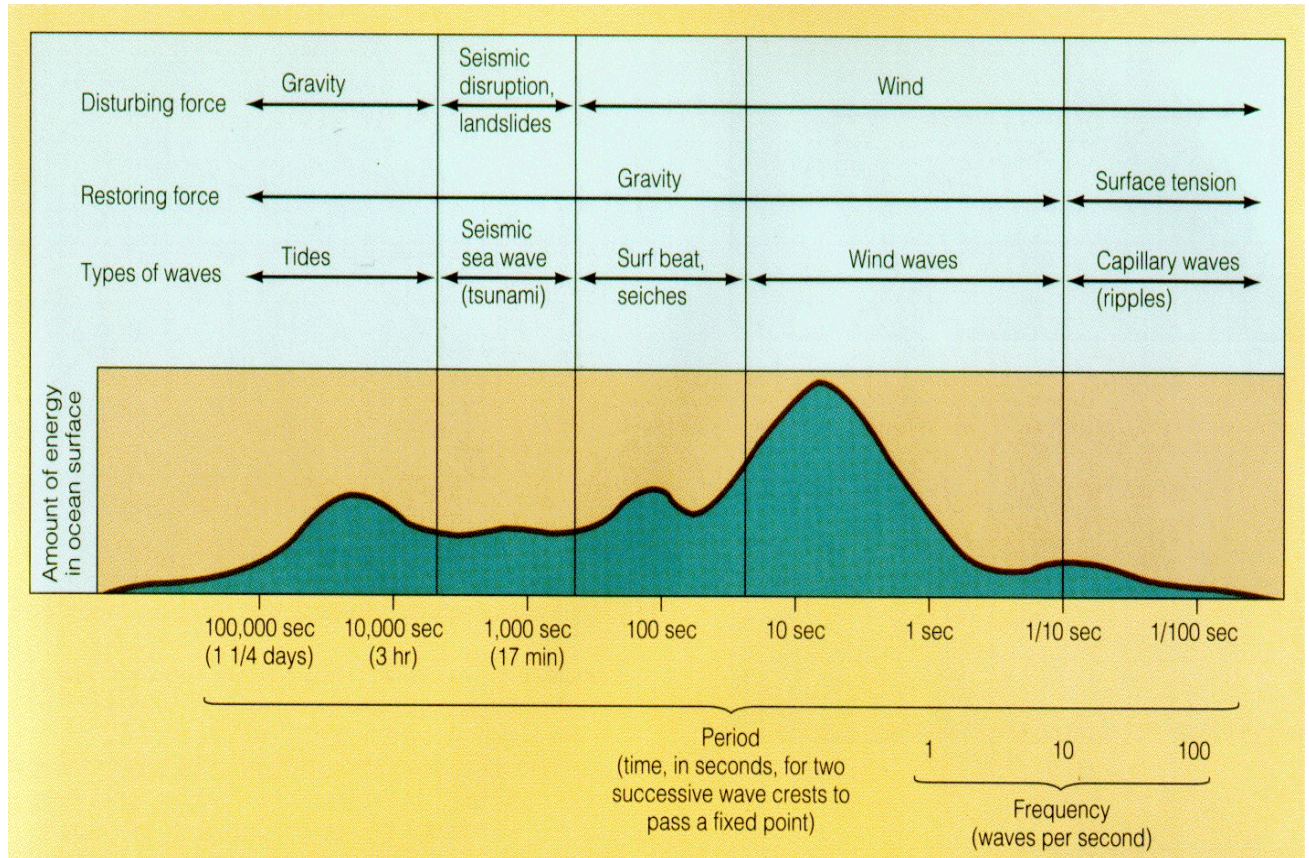


Fig.5.5 Energy spectrum for surface waves

Internal waves

The reason that waves occur at air – ocean interface is the large difference in density. However, in the ocean we find density differences, although much smaller, between the upper layer and the layer below the thermocline. Also on this density interface waves may occur, these are called internal waves. They usually have periods of a few minutes, and may have amplitudes of several meters. These motions may therefore bring nutrients from below the thermocline to the photic zone. These waves are sometimes referred to as “nutrient pump”. An example of observation of internal waves is shown in Fig.5.6.

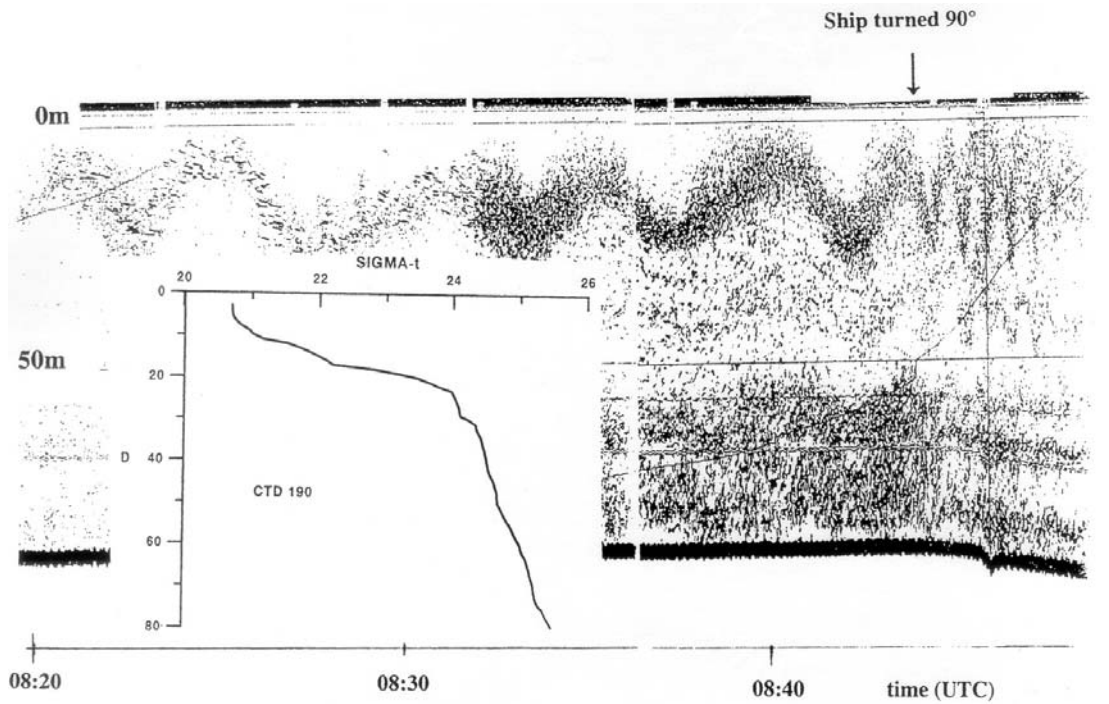


Fig.5.6 Observations of internal waves by an echo-sounder on board R/V “DR. Fridtjof Nansen” outside the Angolan coast in March 1995. The undulating structure between 0 and 30 m depth is due to reflection from phytoplankton in the halocline. After about 20 minutes when the ship was moving parallel to the wave crests, the ship turned left heading into the waves. The corresponding shift in frequency is clearly seen. Inserted the density profile obtained using a CTD.

Langmuir Circulation

These are not waves, as they do not have any phase speed, but they are organized as helical motions with scales of the order of magnitude as the mixed layer, see Fig.5.6. How they are developed is still not clear, but they are often observed as streaks and slicks more or less parallel to the wind direction. As Langmuir circulations also are associated with rather strong vertical velocities, they may also act as a nutrient pump.

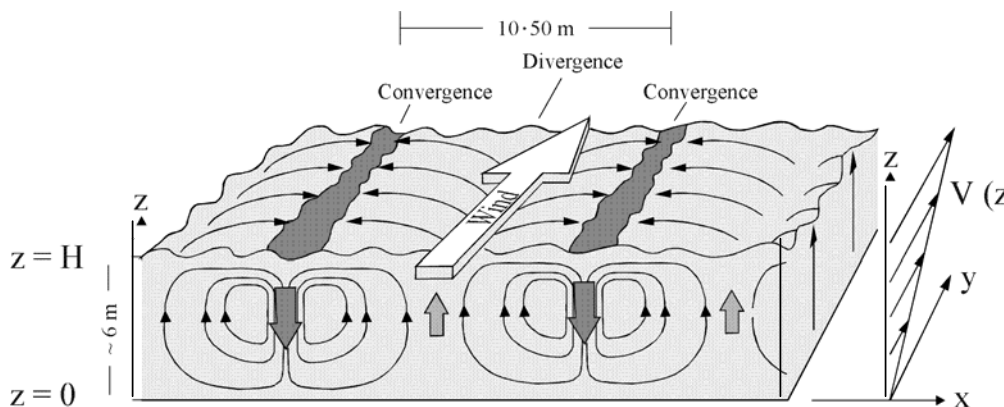


Fig.5.6. Langmuir circulations.

6. TIDES

Newton's law of gravitation says that two masses act on each other with a force proportional to the product of masses and inversely proportional to the square of the distance between the centres of mass, see Fig.6.1.

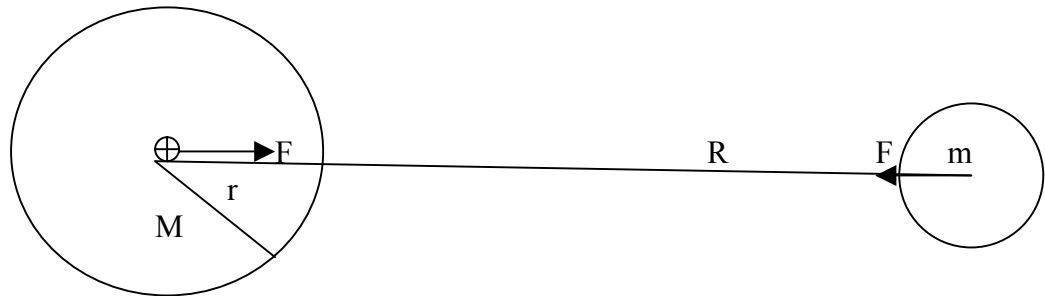


Fig. 6.1. Illustrating Newton 3rd law of gravitation.

The magnitude of the forces F are given by

$$F = G \frac{mM}{R^2}$$

where G is the gravitational constant $G=6.67 \cdot 10^{-11} \text{ N} \cdot \text{m}^2/\text{kg}^2$
 [N=Newton= $\text{kg} \cdot \text{m}/\text{s}^2$]

Now, of course the moon would soon fall down on the earth if it was not orbiting around the globe, and such kept at distance because of the centrifugal force. In fact the moon – earth system orbits around the common centre of mass, (which is situated some place inside the earth). Thus also the earth is subject to a centrifugal force. This explains why the tides have a semi-diurnal, rather a diurnal period, see Fig.6.2.

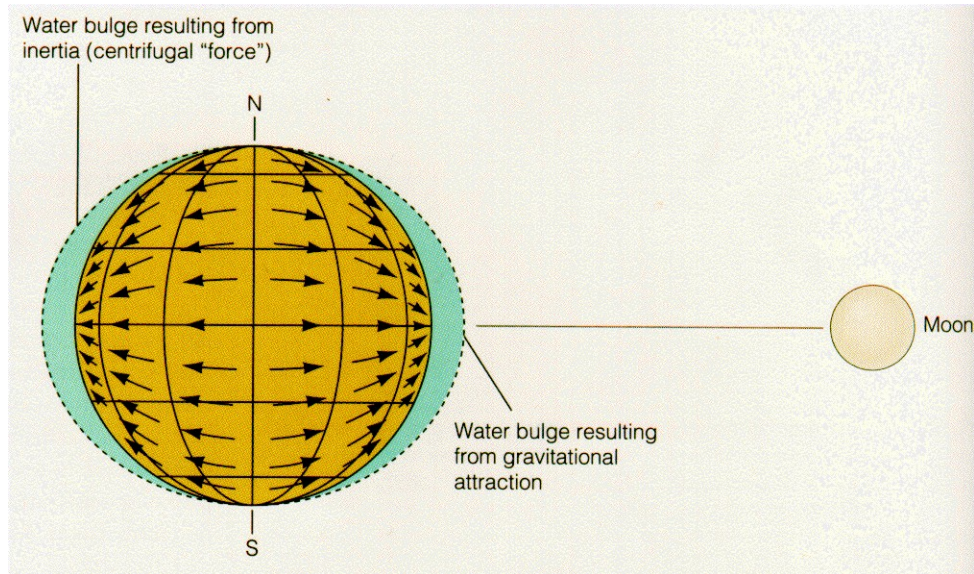


Fig. 6.2 Close to the moon we have a high tide due to the higher gravitational force, and at the opposite side we also have high tide because of the centrifugal force.

If the moon is not situated exactly over the equator, a diurnal period also appears, as illustrated in Fig.6.3. Therefore the tides may also have diurnal periods.

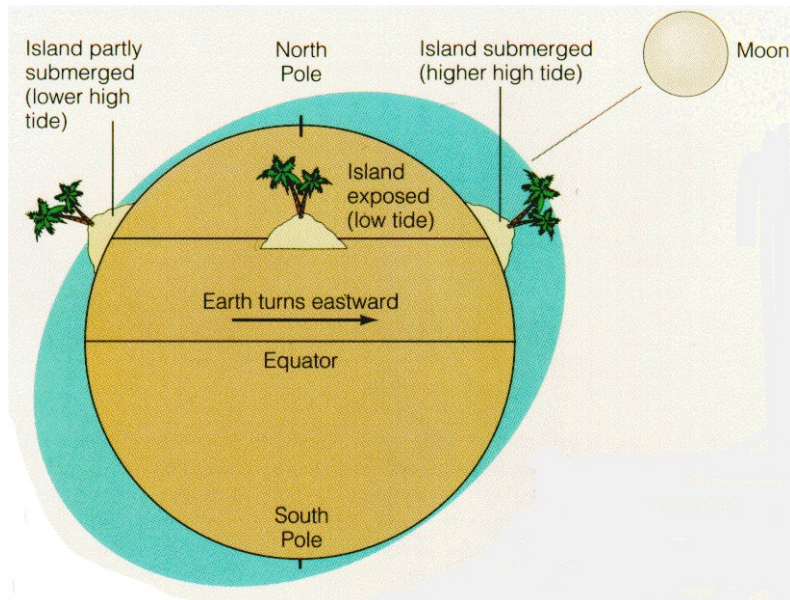


Fig.6.3. Because the rotation plane of the moon is oblique to the earth's axis, diurnal tidal periods also appear.

Therefore the common situation is a mixture of diurnal and semi-diurnal tides, see Fig.6.4.

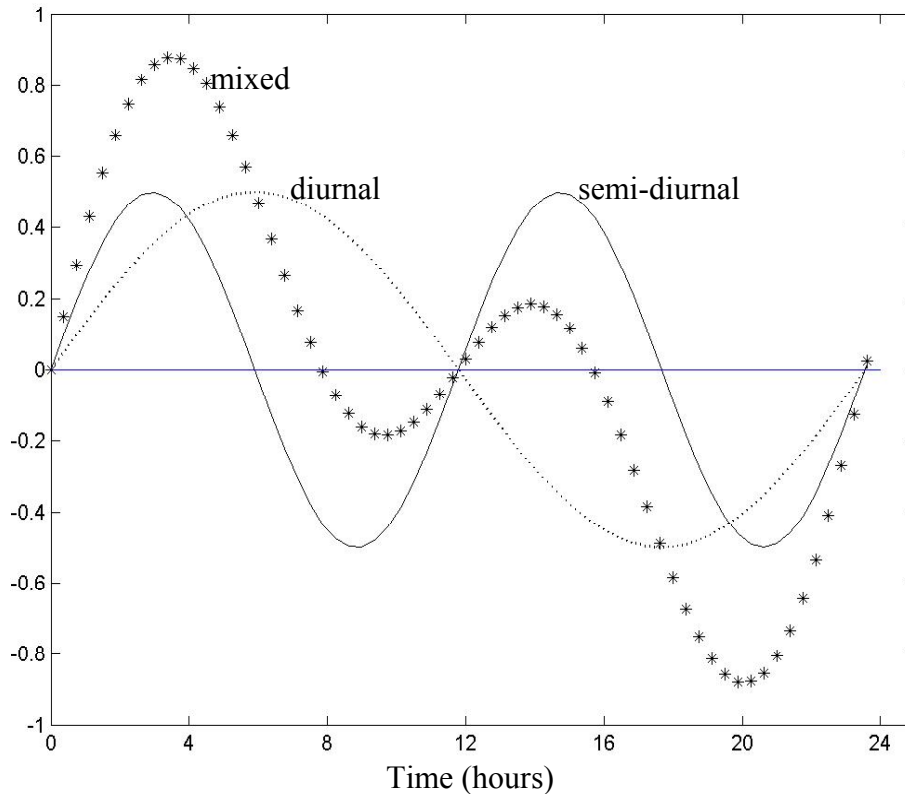
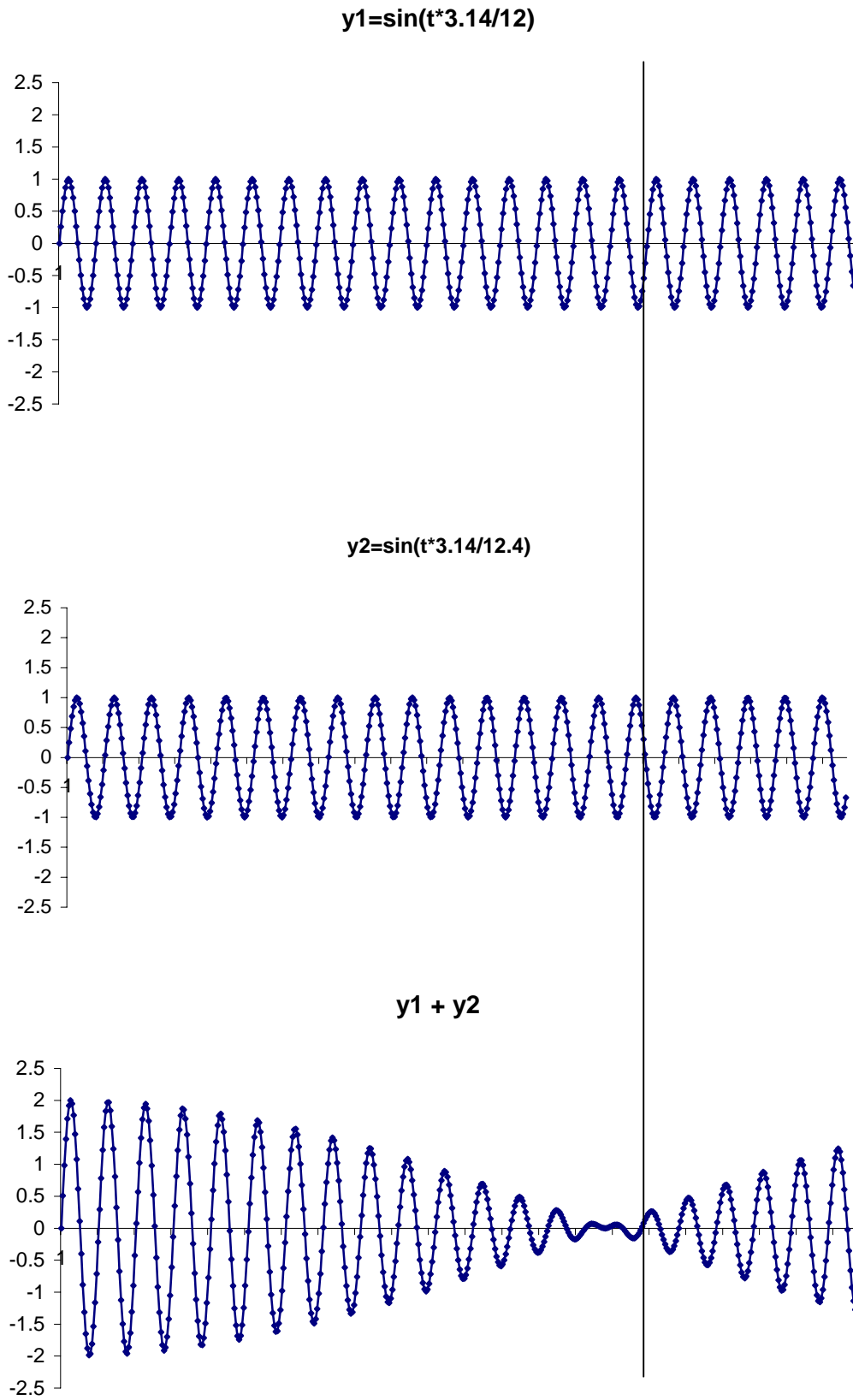


Fig.6.4 Illustrating how a diurnal and semi-diurnal period adds up to a mixed tide.

Now we will consider the tides caused by the attraction from the sun. Of course we again have to apply Newton's 3rd law of gravitation. The mass of the sun is much bigger than the mass of the moon, but because the distance is so enormous it turns out that the tides caused by the sun has amplitudes about half of those caused by the moon. It is important to notice that the tidal periods caused by the sun and the moon are not exactly equal. The diurnal solar period, i.e., the time it takes from the sun is in zenith until it is back in zenith, is of course 24 hours. But because the moon is orbiting around the earth, it will take a little longer before the moon is back in zenith, namely about 24.84 hours. If we simply add two waves with periods 12 hours and 12.42 hours, the superposition principle shows that the result is a wave which changes from maximum amplitude to minimum amplitude in about one week, see Fig.6.5



Spring tide

Neap tide

Fig.6.6. Adding solar and lunar semi-diurnal tides.

These bi-weekly extremes are called spring and neap tides. Why it is like that is easy to understand if we consider the relative position of the sun and the moon, see Fig.6.7.

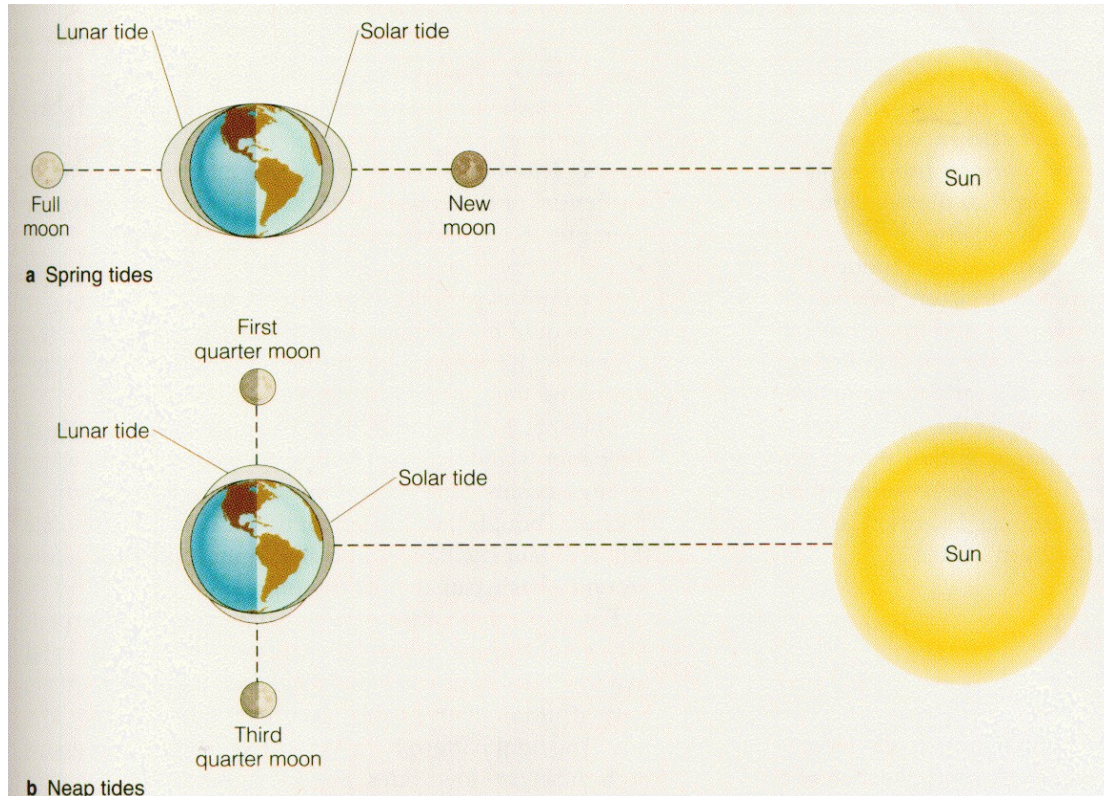


Fig.6.7. The solar and lunar tides interact to cause spring and neap.

The tidal structure is of course not so simple as we have indicated above. When large water masses are moved from one place to another via the tidal currents, these movements will be disturbed by topography and continents. Therefore maps have been constructed showing how the tidal amplitudes vary over the globe, so called co-tidal maps, see Fig.6.8.

Some of the major tidal constituents are shown in Table 6.1

Name of tidal component	Symbol	Period in hours	Relative size ($M_2 = 100$)
Principal lunar, semidiurnal	M_2	12.42	100
Principal solar, semidiurnal	S_2	12.00	47
Luni-solar, diurnal	K_1 -solar	23.93	58
Principal lunar, diurnal	O_1	25.82	42

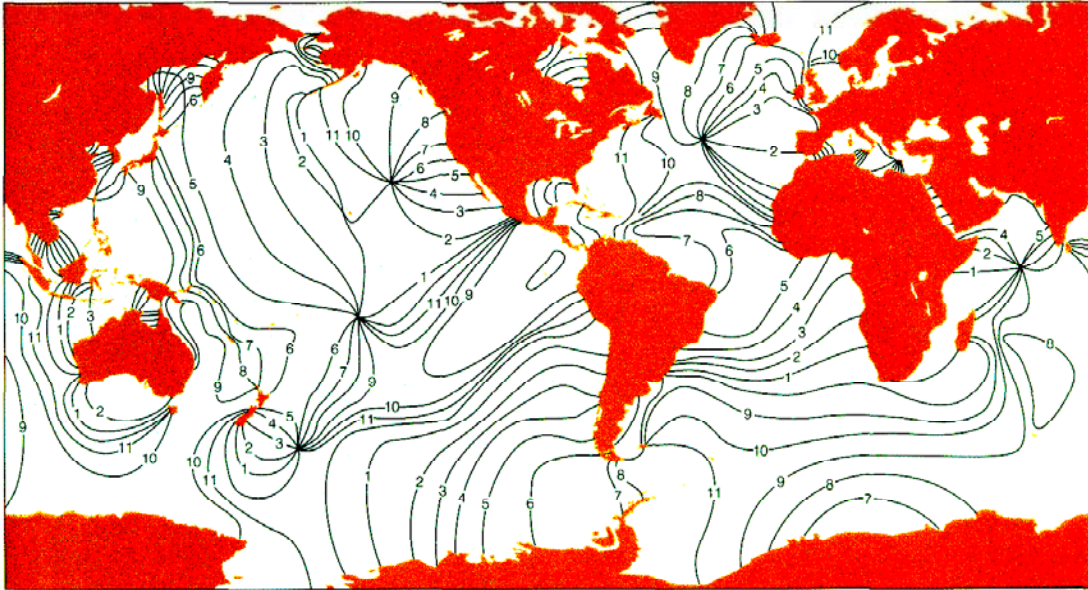


Fig.6.8. Co.tidal map of M_2 (lunar semidiurnal). The lines are called the co-tidal lines, and the numbers indicate the time of high water (in hours) after the Moon has passed the Greenwich meridian. The positions where the co-tidal lines meet, are called amphidromic points. At an amphidromic point the corresponding tidal amplitude is zero. The amplitude will in general increase with distance from an amphidromic point.

7. THE BENGUELA SYSTEM

The Benguela is one of four major coastal upwelling systems of the world ocean. The Benguela is unique, compared to the California, Peru and North West Africa, that it is bounded at both equatorward and poleward ends by warm water regimes: To the north we find the tropical waters of Angola, and in the south the warm Agulhas current is dominating.

Fig.7.1 shows the bathymetry of the region. The shallow coastal shelf, defined by the 200 meter isobath, is typical ~100km wide. Here most of the rich fisheries in the region take place. The deep basins, Angola Basin and Cape Basin are more than 5km deep. They are separated by the Walvis Ridge. In the south we find the Agulhas Ridge.

The average current picture is shown in Fig.7.2. Some distance off shore we find the main branch of the Benguela current with a transport of about 15 Sv (1 Sv = 1 Sverdrup = $10^6 \text{ m}^3/\text{s}$).

Closer to the shore the surface current is in general directed northwards until it meets the south-flowing Angola Current near the Cunene River. Here we find a very marked front in the SST, called the Angola Benguela Front, see Fig.7.3b.

Between Lüderitz and Cape Columbine we find a near-shore surface counter-current flowing southwards.

Below the surface on the whole shelf region we usually find a poleward undercurrent.

Typical water masses at the shelf are illustrated by vertical sections of T, S and O₂ at 18°S, see Fig.7.4a. Note that although the section was taken in March, near the peak summer, the surface temperature is only slightly above 20°C. The salinity ranges between 35.2 and 35.6. Note the oxygen-depleted layer near the bottom. This feature is occasionally found on the Namibian Shelf, and is typical near Walvis Bay. The reason for this oxygen minimum layer is still disputed, but it seems to be a combination of local biological activity and low oxygen water flowing in from the north.

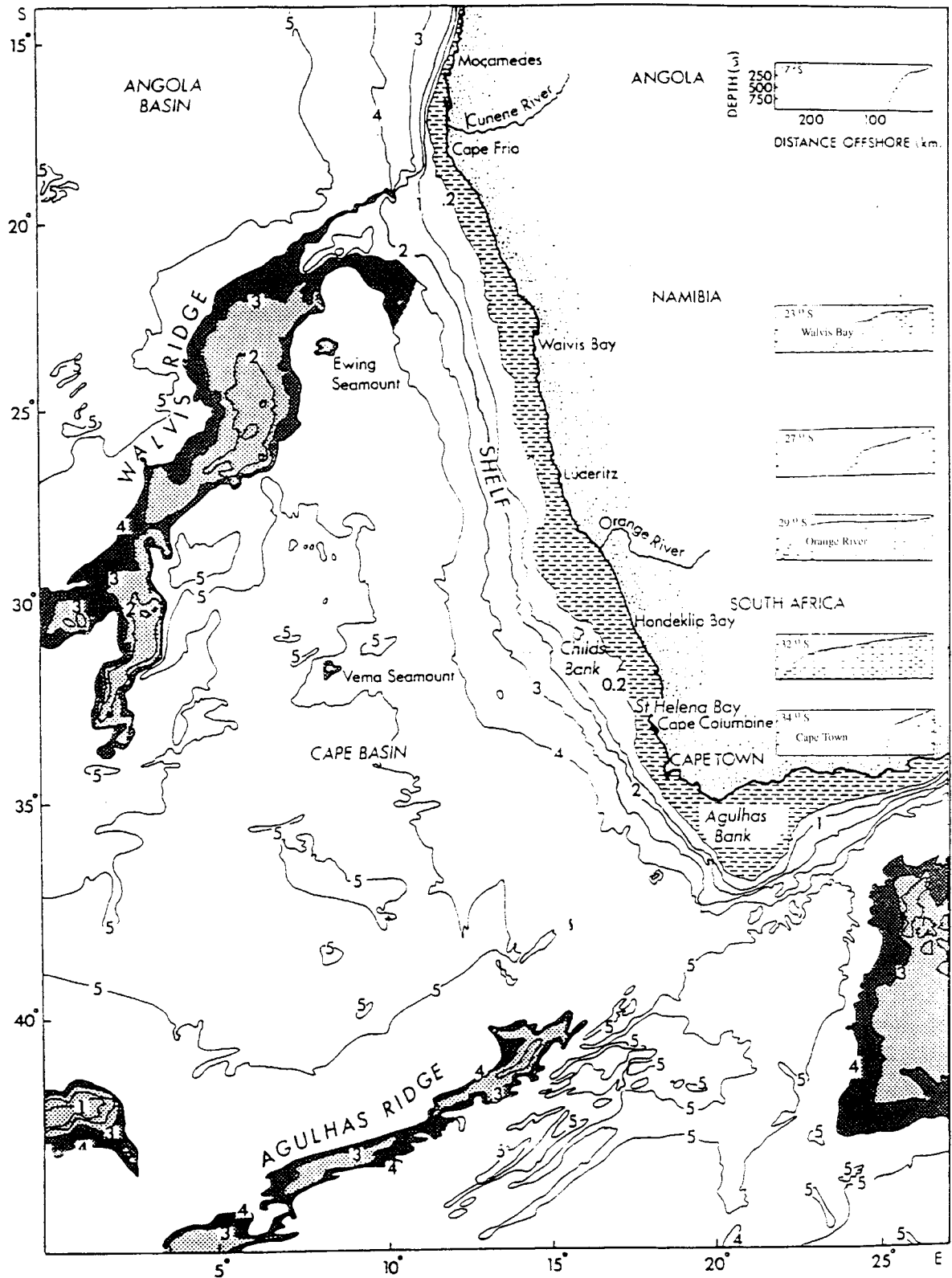


Fig.7.1 Bathymetry of the southeast Atlantic Ocean and profiles of the shelf at selected latitudes.

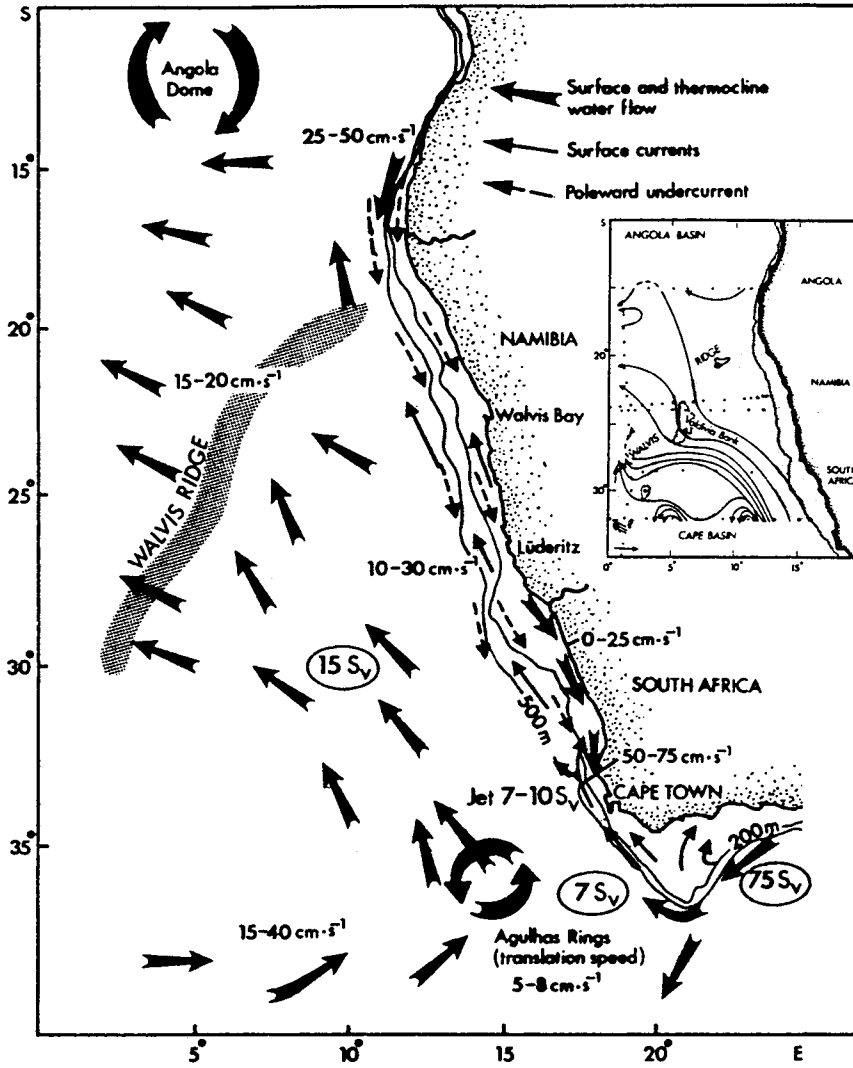


Fig.7.2. Schematic flow fields of surface and thermocline water. Current speeds refer to surface values. Transports (circled) refer to total transport above 1500m.

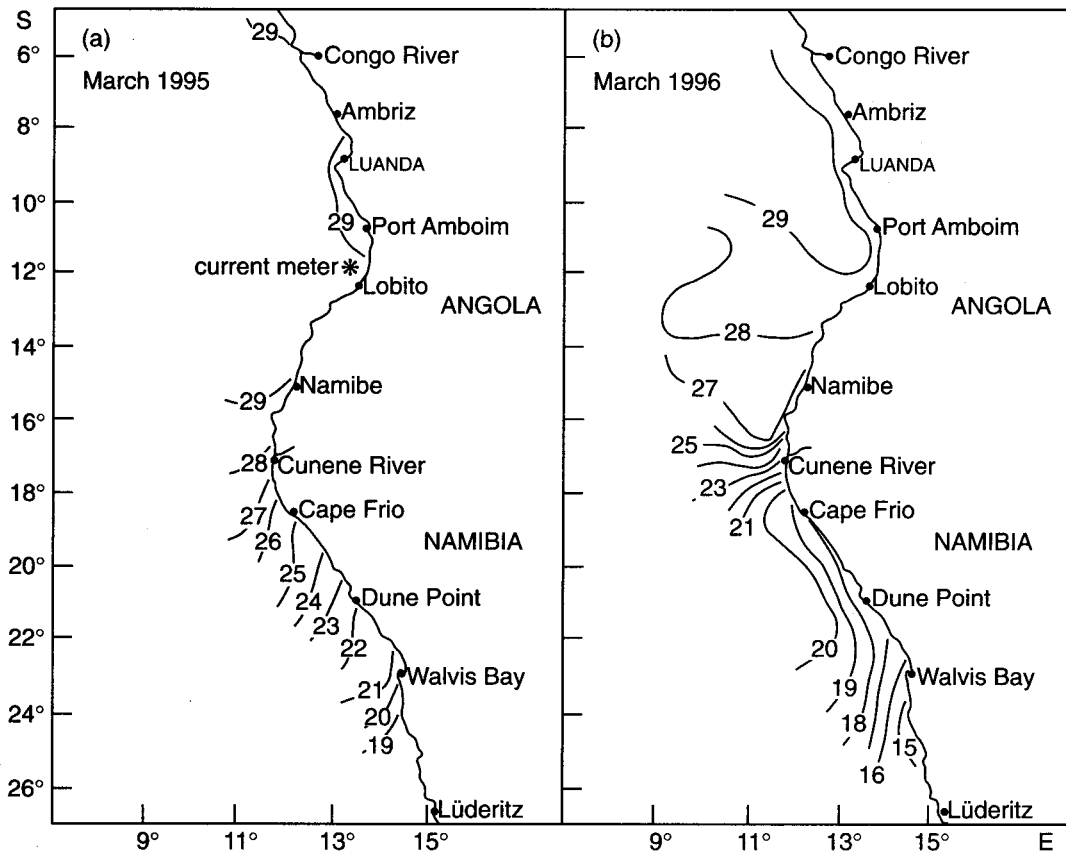


Fig. 7.3 SST ($^{\circ}\text{C}$) in March for a) A Benguela *Niño* year (1995), and b) A normal year (1996)

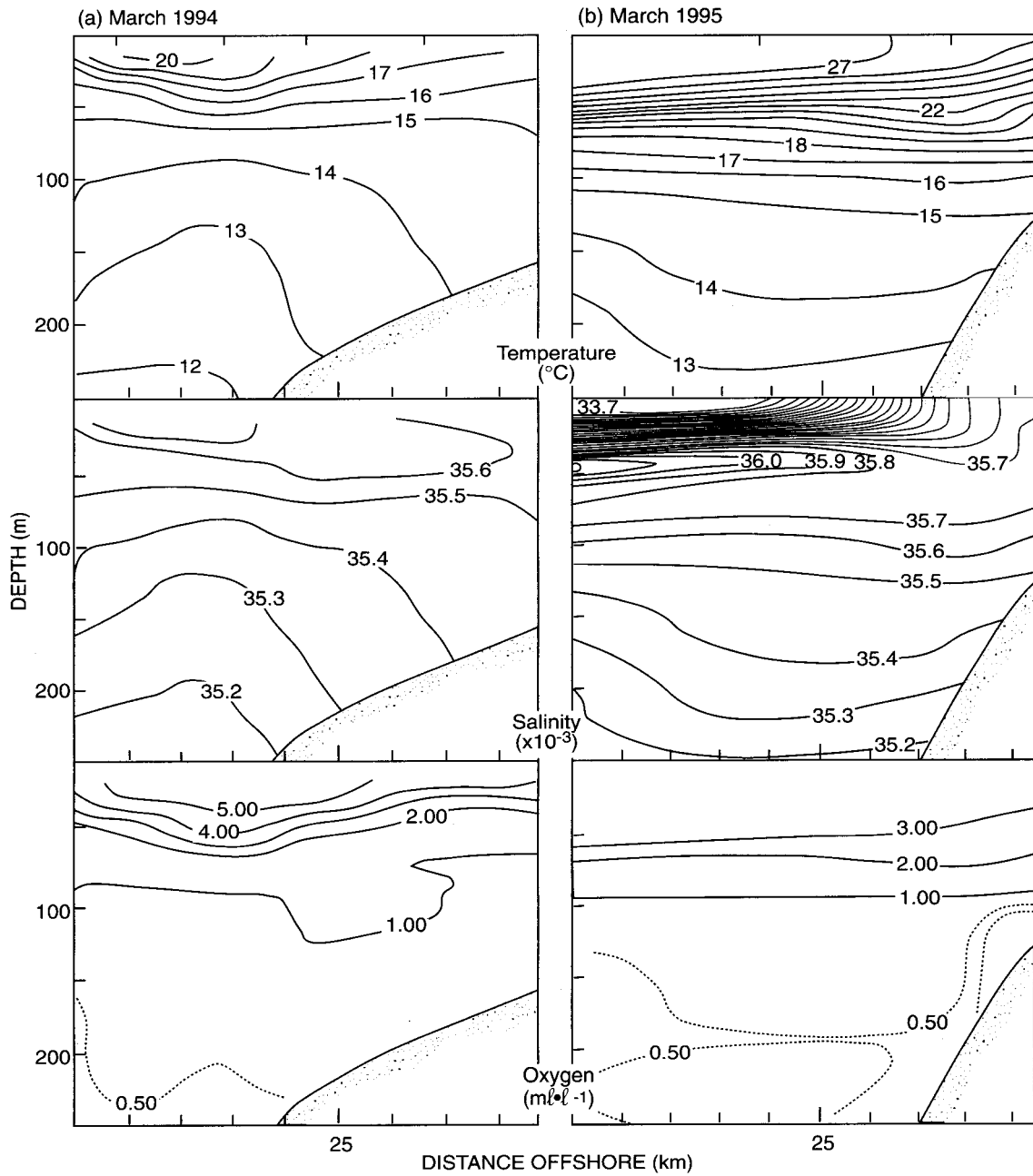


Fig.7.4. Vertical sections of temperature, salinity and oxygen near 18°S in a) March 1994 and b) March 1995.

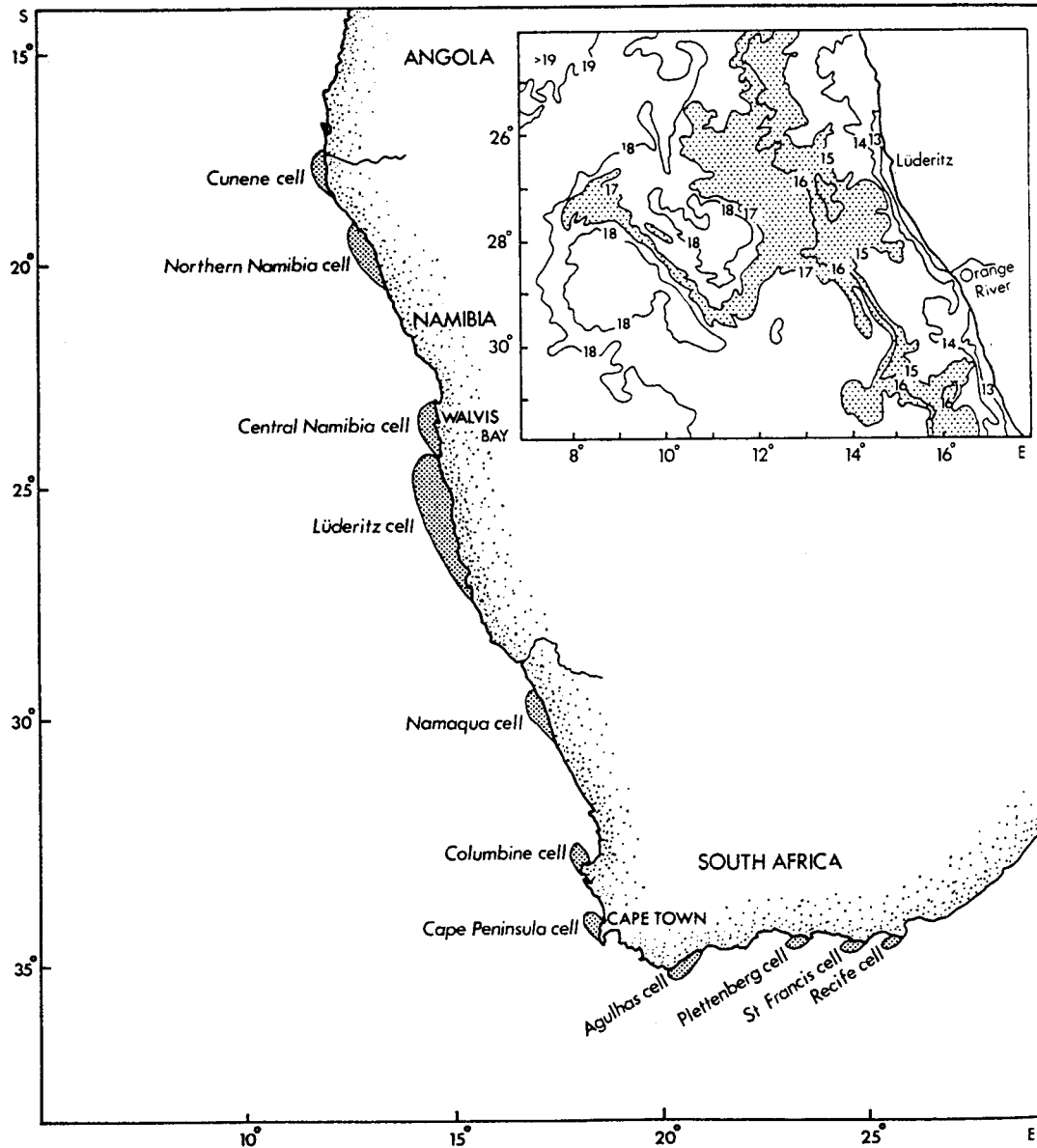


Fig.7.5. Areas along the Namibian coast where the upwelling is known to be intense are defined as upwelling cells.

Upwelling cells.

The upwelling is not equally strong along the whole coast. Some regions where the upwelling is more intense, are defined as upwelling cells, see Fig.7.5. The position of the upwelling cells may be determined by the wind conditions. (Why?). In Fig.7.6 the wind force acting at the ocean surface as a function of latitude and time is shown. Fig.7.6 indicate that the upwelling intensity should vary both along the coast and with season.

Exercise: Compare Figs. 7.5 and 7.6, and discuss if the position of the upwelling cells may be explained by the wind forcing.

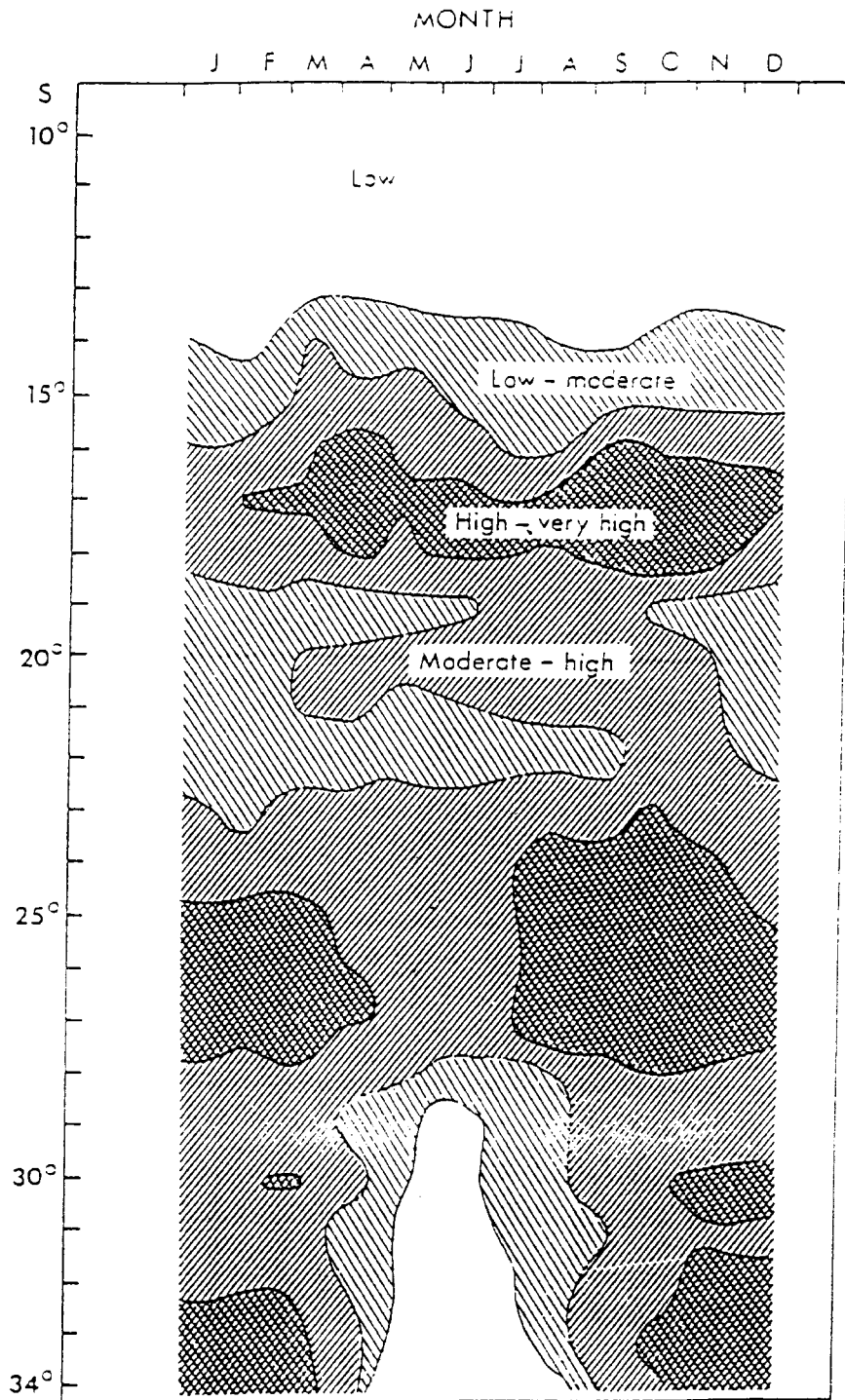


Fig. 7.6. Monthly wind stress along the west coast of Africa, centres 40 nautical miles offshore at whole degrees latitude (after Boyd 1987)

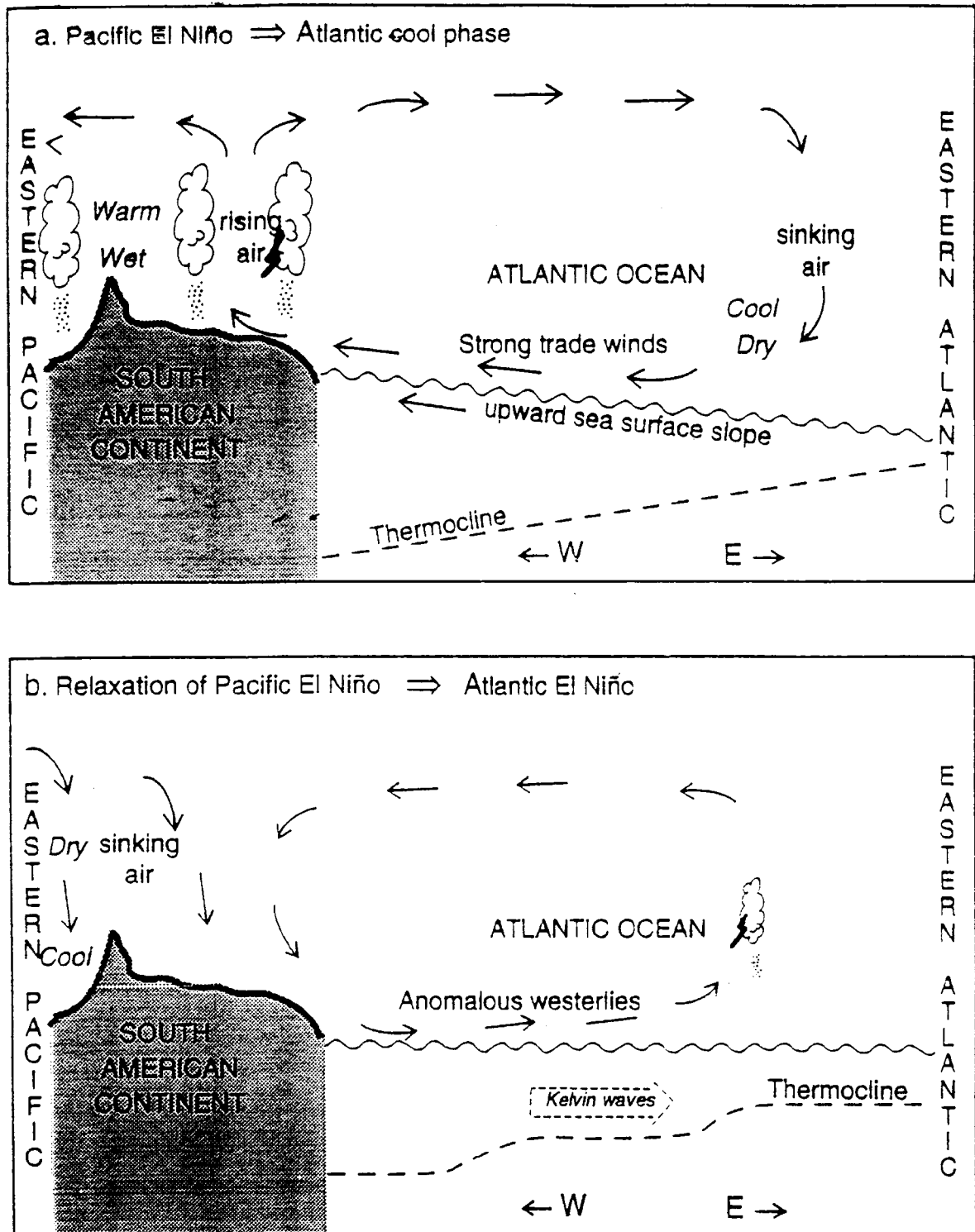


Fig. 7.7 Diagram illustrating a hypothetical “tropical” ENSO teleconnection between the Pacific and Atlantic. The figure represents the vertical plane along the equator. a) During a Pacific ENSO phase, the strengthened Atlantic trade winds, due to anomalous expansion of the zone of rising air over northern South America, build up anomalous trans-Atlantic slopes in the sea surface and thermocline along the equator, and associated anomalous surface temperature contrasts. This sets up the stage b) an El Niño – like collapse of the intensified Atlantic system the following year.

Benguela Niños

The well-known Pacific *El Niño* is manifest as an episodic warming of coastal waters off Peru. This warming is related to a shift in the wind regime in the Pacific, reducing or reversing the winds towards the west, which normally keep the warm water off shore. The *El Niños* are known to have a detrimental influence on the fisheries. It is less known that the Pacific *El Niño* has a counterpart in the South-East Atlantic. These are less frequent and intense than the Pacific events. They occur about once every 10 years, and are known as *Benguela Niños*. Even the *Benguela Niños* are believed to be related to the shift in the wind patterns, see Fig.7.7.

During the last *Benguela Niño*, which occurred in 1995, the temperature at the surface along the Namibian coast was anomalously warm, see Fig.7.3a. Comparing with Fig.7.3b we note that the Angola-Benguela Front is smeared out, and the temperature at Walvis Bay, for instance, was about 5 degrees above normal. The surface heating is also revealed in the vertical section (Fig.7.4b). Also note that the *Benguela Niño* in 1995 was associated with a low salinity surface layer (Fig.7.4b).

Fig.7.8 shows that the 1995 *Benguela Niño* forced the sardine southward.

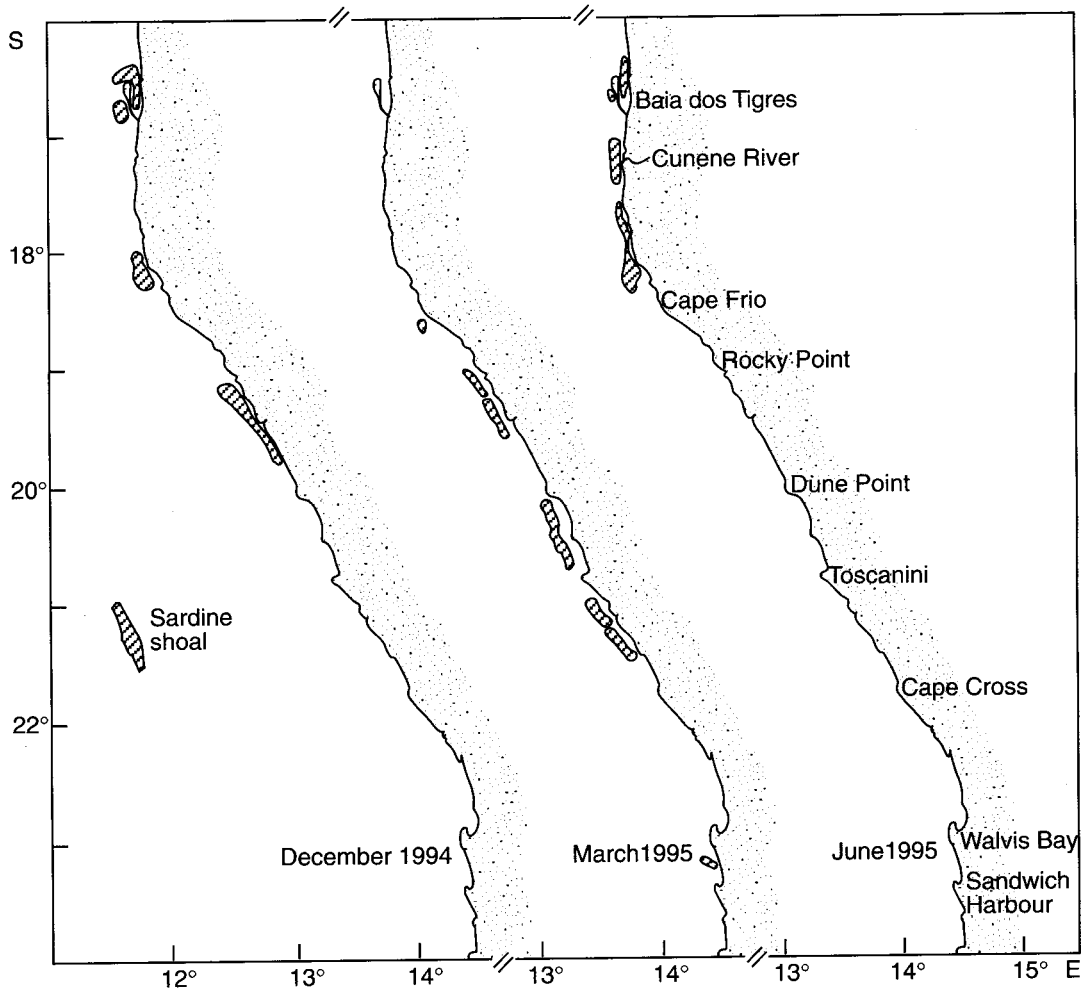


Fig.7.8 Distribution of sardine in northern Namibia and southern Angola in December 1994, March 1995 and June 1995, based on acoustic surveys

8. ENVIRONMENTAL CONDITIONS AND FISHERIES.

This is a field of very active research. Therefore, do not consider what is said below as the final truth. Rather, I have tried to give you the “state of the art” (2001).

Global variations and climate

Several of the important fisheries in the world show a decadal variation, see Fig 8.1. There seems to be a synchronous variation of sardine abundance at various sites in the Pacific Ocean, Fig 8.1 a. On the other hand, another small pelagic in the Pacific, the Peruvian anchoveta (Fig.8.1b), seems to be out of phase with the sardine, but in phase with the South African pilchard. It is very difficult to find a pure biological explanation of this behaviour. The sardine stocks in the various areas of the Pacific are not related at all, and the fishing pressure at the various sites have been different, and did not vary in a manner that have resemblance to Fig.8.1. It is interesting to note that the Peruvian anchoveta was building up in the 60's when the abundance of the South American sardine was low. One has tried to explain this mutual exclusion behaviour by predator – prey relationships, but with no success. And why should Peruvian anchoveta and South African pilchard behave in the same manner?

It seems that we have to look at the environmental conditions to get an answer. The worlds' climate also show decadal variations, see Fig. 8.1 c, which show the fluctuations of the SST's in the central Pacific. Such decadal climatic variations do not vary simultaneously in the world oceans, but seem to be passed from one ocean to the next via the atmosphere, i.e. so-called teleconnections. The most famous example of such tele-connection is the El Niño, which has its counterpart in the Atlantic, see Fig. 7.7.

Although we have represented the climatic variability in the ocean by temperature (SST) in Fig. 8.1c, it does not mean that it is the temperature *per se* which is of importance. The temperature of the ocean is a result of variations in the atmosphere, (clouds, radiation etc.), and particularly changes in the wind-systems. Although temperature indeed have an important influence of various biological processes, (examples are given below), it may well be that changes in the wind systems has more direct influence on the fisheries. For example, in the case of El Niño, it is well known that the ocean temperature becomes anomalously warm. But this warming is a result of changes in the wind systems, see Fig. 7.7. The SST increases mainly because the thermocline depth increases. However, beside the obvious direct influence of temperature on the biological processes and fish behaviour, the fact that the nutrient

rich water below the thermocline not is as easy available for the photic zone may be equally important.

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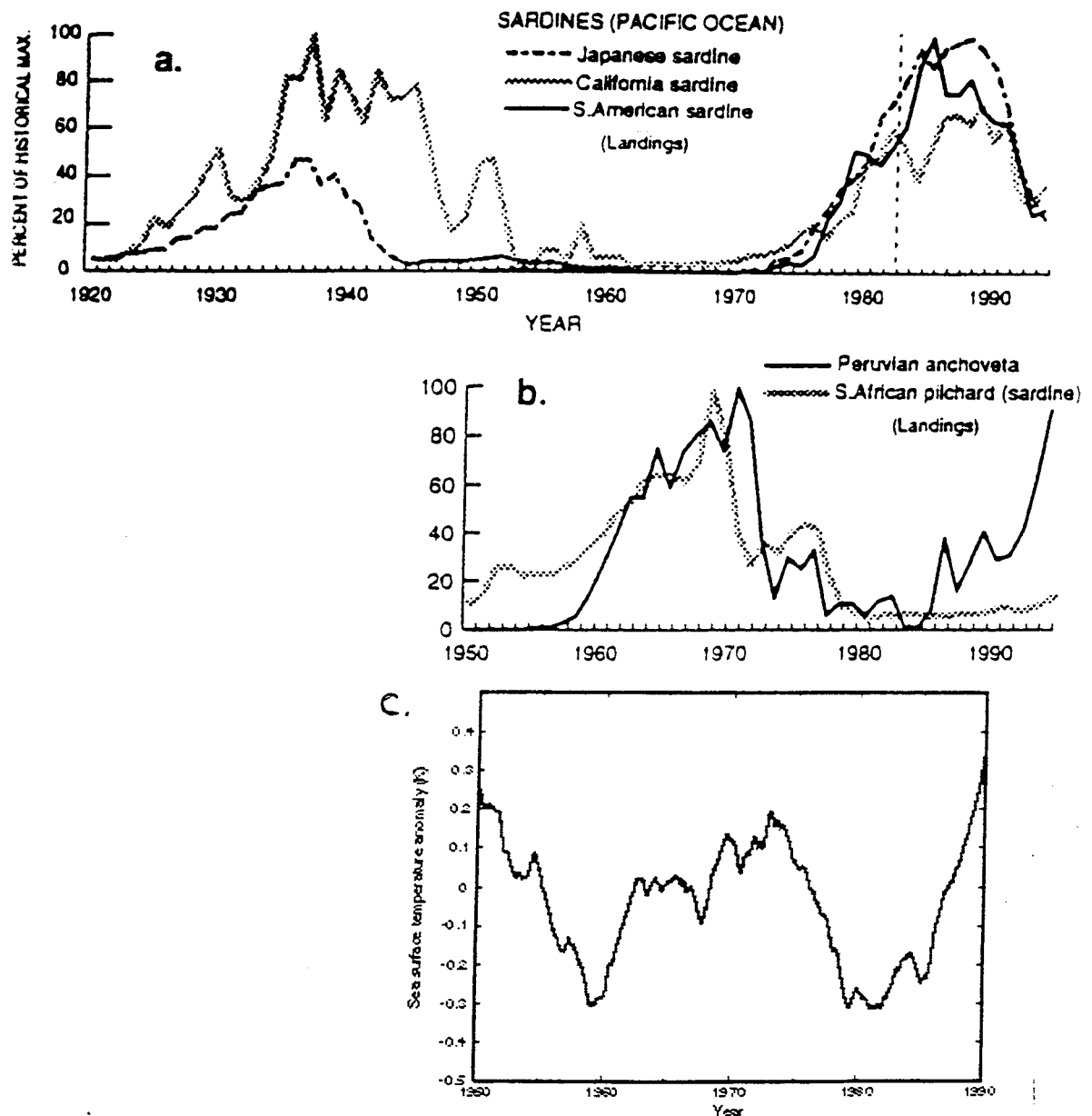


Figure 8.1. Time variations of catch given as % of historical maximum of a) Sardines (Pacific ocean), b) Peruvian anchoveta and South African pilchard, and c) SST anomaly in central Pacific ($25^{\circ}\text{N} - 40^{\circ}\text{N}$, $170^{\circ}\text{E} - 160^{\circ}\text{W}$) (from a model simulation).

Micro-scale. Environmental window

The most critical phase in life cycle of a fish seems to be the first feeding of larvae. (The typical state of fish larvae (all species) is that it is dead). The food may be made be available for the larvae in two ways: 1) The concentration of the food is so large that the larvae may obtain it without using much energy, or 2) The food must be brought to the larvae. Recent investigations show that turbulence may be a key factor in this process. Increased turbulence may increase the encounter-rate between food particles and larvae, but too vigorous turbulence may disperse the food particles. So there seems to be an “environmental window” where the conditions are optimal for larval survival, see Fig 8.2.

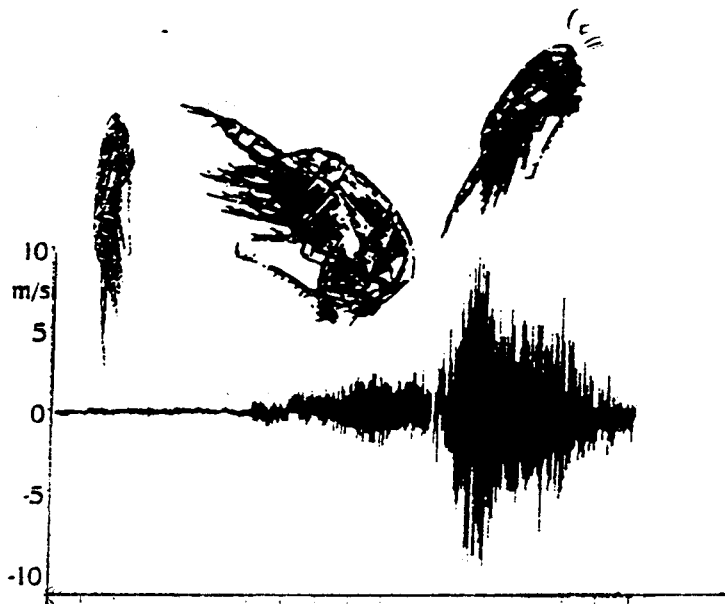
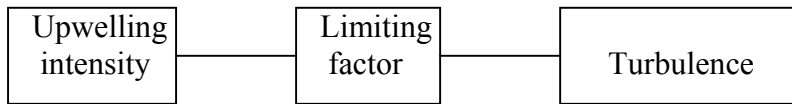


Fig 8.2a. Suggested correlation between “appetite and well-being of copepods and turbulence intensity. Low intensity: starving copepods; encounter-rate too small. Medium intensity: well being copepods. Strong intensity, stressed copepods; food particles pass faster than the copepods can attack

Another example of an environmental window at the upper trophic level is shown in Fig. 8.3, where catch of albacore is compared with SST. One should remember, however, that there are strong relationships between ocean temperatures and some biological processes, see Fig. 8.4.



Recruitment

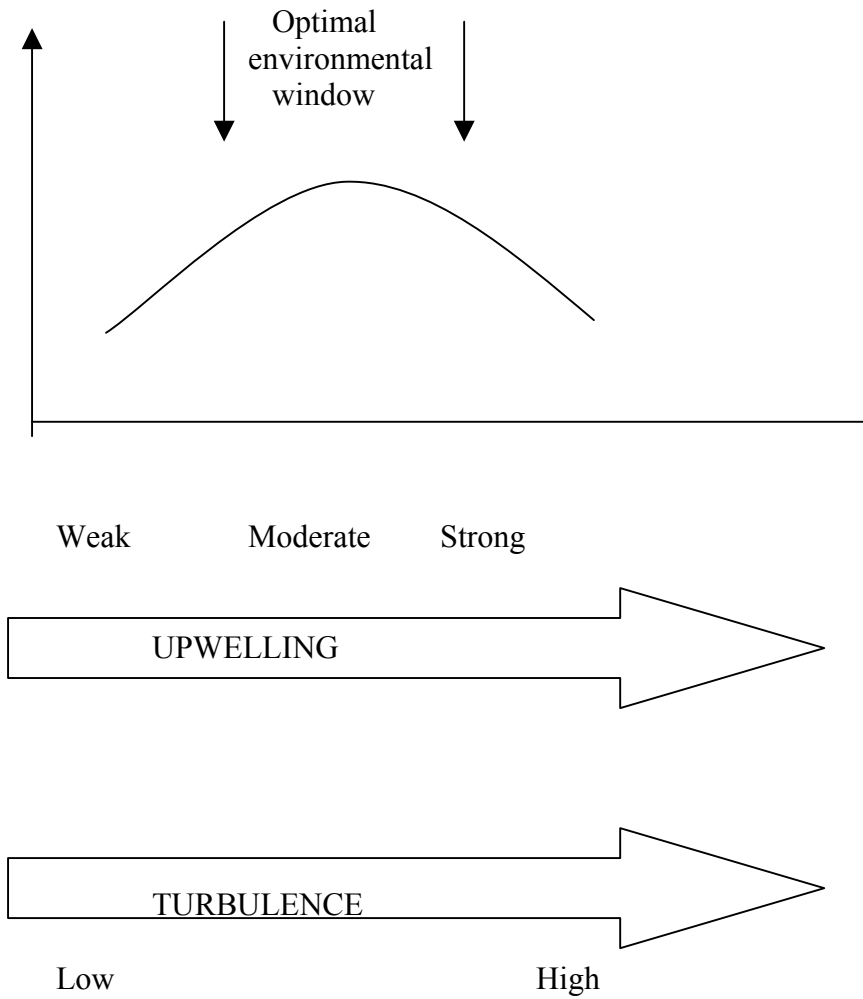


Fig. 8.2 b Schematic of an environmental window

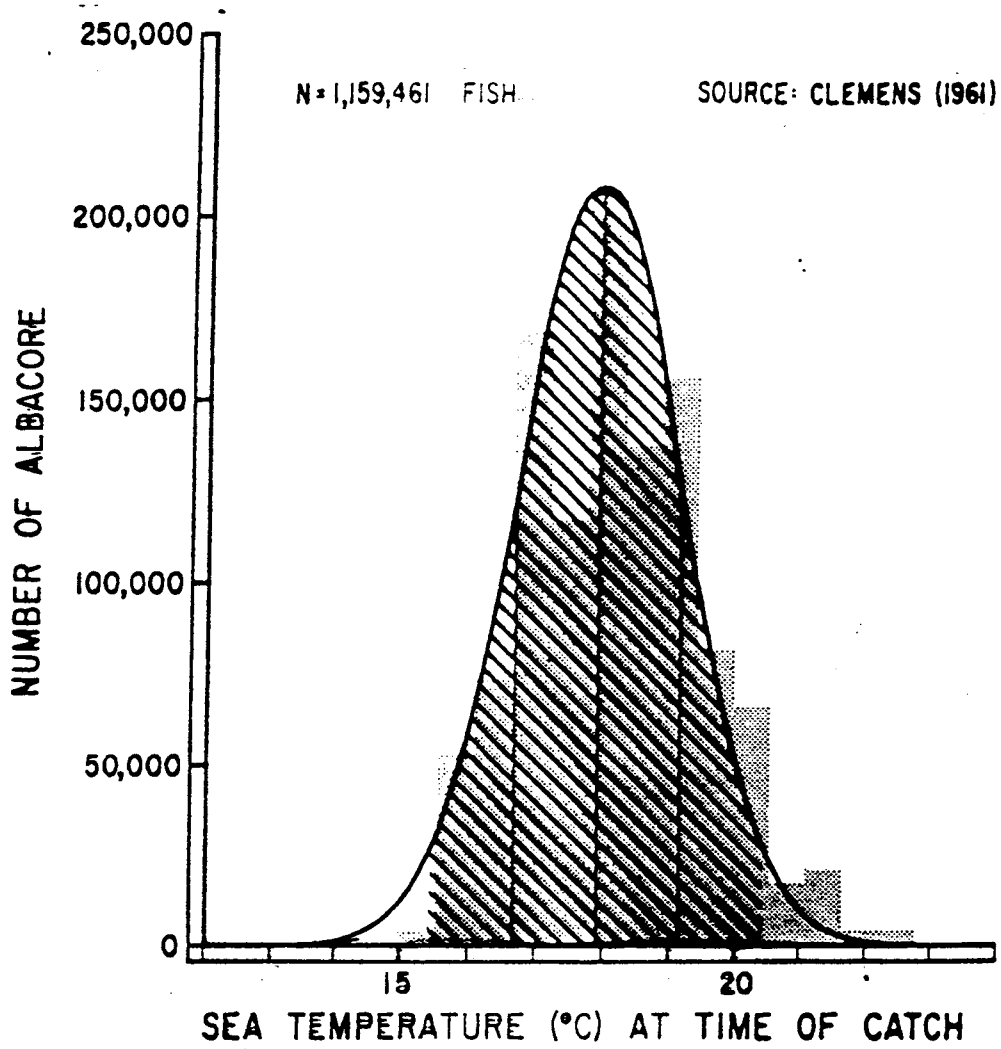


Figure 8.3 The relation of SST (°C) and the albacore catch. Hatched area includes 95% of total catch

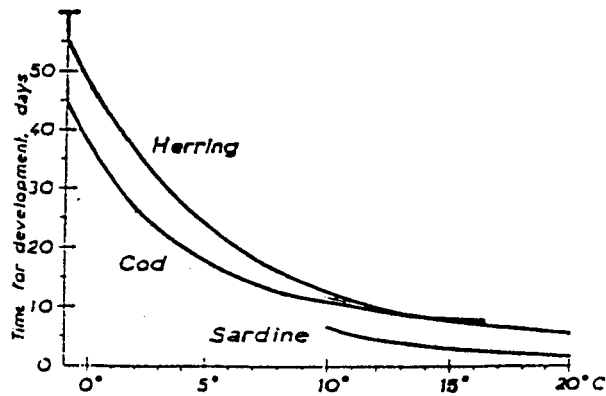


Fig. 3-3. Incubation periods of herring, cod and sardine eggs.

Figure 8.4. Incubation periods for some fish eggs as a function of temperature.

Ocean triads

Comparative studies of geographical climatology of fish reproductive habitat have tended to identify a “fundamental triad”, also called “Bakun triad” (Bakun 1996) of three major classes of processes that combine to yield favourable reproductive habitat for coastal pelagic fishes and also many other types of fishes:

- 1) *enrichment* (upwelling, mixing, etc)
- 2) *concentration* (convergence, frontal formation, water column stability)
- 3) *retention* (drift toward or retention within appropriate habitat)

The importance of *enrichment* is quite intuitive, and moreover, tends to be clearly evident in geographical patterns of fish abundance. Most of the ocean is quite unproductive because of scarcity of certain essential plant nutrients, which are trapped below the photosynthetic layers by ocean stratification. Areas where nutrients are supplied to the illuminated ocean surface layers by upwelling or mixing tend to be highly productive and to contain large, prolific fish populations. The coastal upwelling systems covers only about 0.2% of the world ocean, yet more than 50% of the world ocean fish resources are caught in these areas.

Perhaps less widely appreciated is the importance of *concentration* processes. The importance of high concentration of food particles being available for larvae at their first feeding stage was mentioned above. An important type of concentration processes is indicated in Fig. 8.5, and occurs at frontal interfaces between waters of differing densities, for instance between oceanic surface waters and recent upwelled waters, at the edge of a river plume, etc. The buoyancy driven flows which supply this mixed water formation are directed from each of the water types toward the interface,

resulting in a zone of convergence that acts to sustain the distinct frontal character of the boundary. Tiny, weakly-swimming organisms that may be unable to resist being passively swept along in the horizontal ocean flow field may be well able to control their depth level in the much less energetic field of vertical motion in the ocean. Such organisms will thus accumulate in the slowly sinking waters of the convergent zone and the distribution of food particles would tend to be highly concentrated (Fig. 8.5).

The third factor in the “triad” is *retention*. Life cycles of marine organisms tend to include at least one stage of passive larval drift. Thus, in a dispersive fluid medium, loss of early life stages from the population habitat may represent serious wastage of reproductive resources. Consequently, fish populations tend to spawn in locations and seasons that minimise such losses. It is notable that upwelling centres, with their conditions of strong offshore transport (tending to cause loss of larvae from the habitat) and of intense turbulent mixing (tending to prevent maintenance of fine scale patches or strata of food particles), are avoided as spawning habitats in spite of the very strong enrichment that occurs there.

On broad continental shelves with upwelling favourable winds a double cell upwelling system may occur, which seems to be optimal for biological production as both enrichment, concentration and retention processes are taking place, see Fig. 8.6.

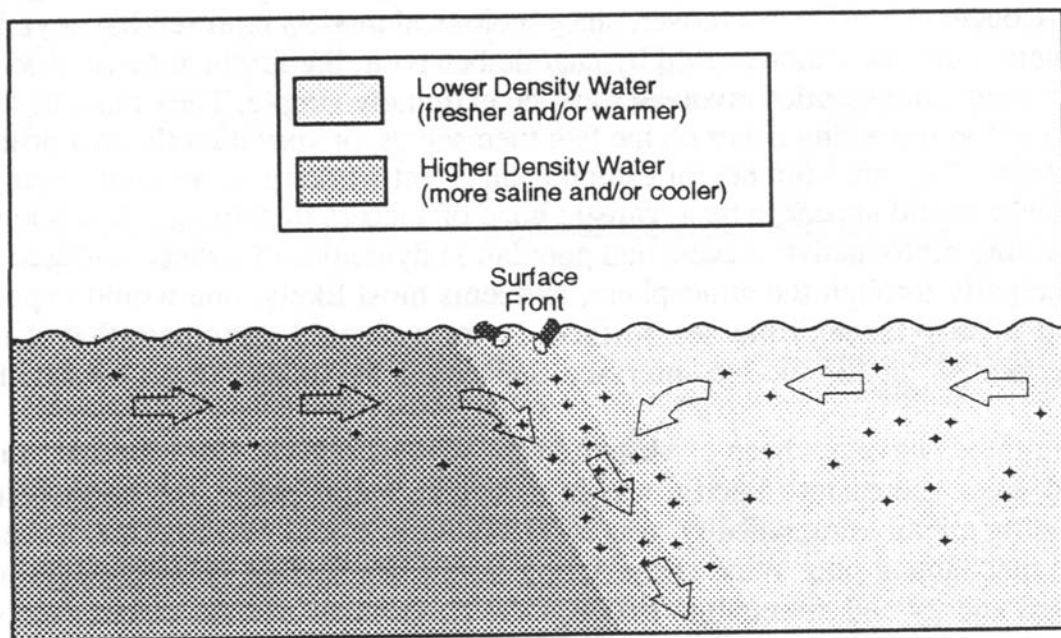


Fig.8.5 Schematic diagram of a front between waters of different density. Arrows indicate density-driven flows associated with the front. “Particle” symbols indicate planktonic organisms capable of resisting vertical displacement. (Scales are distorted: vertical scale greatly expanded relative to horizontal; particles greatly magnified; surface waves not to scale, etc.)

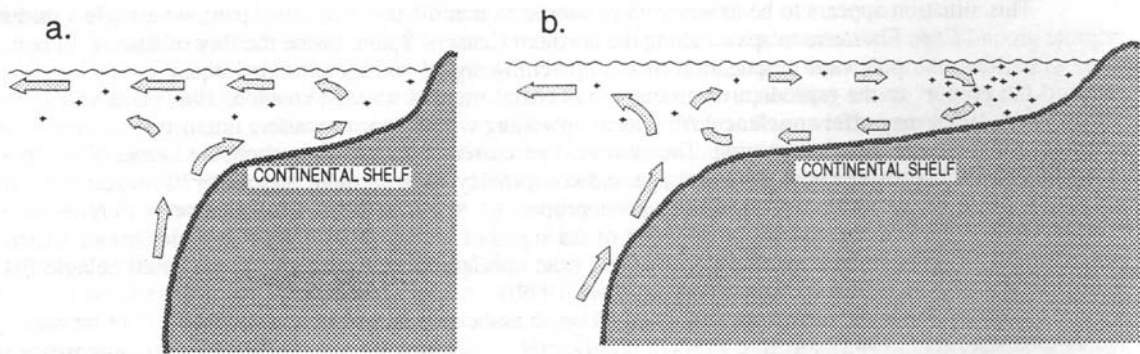


Fig.8.6 Schematic profiles of upwelling circulation patterns (modified from Roy, 1997) and associated distributions of planktonic organisms. a) Normal upwelling circulation b) "Double celled" upwelling circulation, where enrichment occurs at the shelf break. Enriched surface water is carried shoreward to a zone of sinking water next to a coast, which produces concentration and retention.

Other environmental – fish relationships

River run-off

River water may be rich in nutrients (enrichment). Also larvae may be transported by river flow towards nursery grounds (retention). Also there will be a front between the river water and the oceanic surface water where the river runs into the ocean (concentration). Thus all 3 factors in the "triad" may be fulfilled in connection with river run-off.

An example of the obvious connection between river run-off and shrimp catch outside the mouth of the Zambezi River is shown in Fig.8.7.

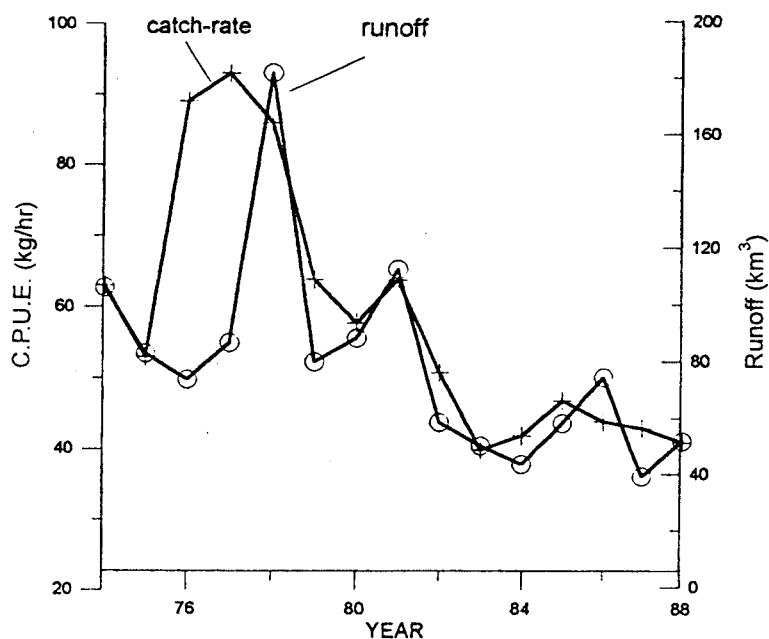


Fig.8.7. Yearly values of Zambezi River run-off and catch per unit effort of shrimps on the shallow shelf outside the Zambezi River mouth in Mozambique.

Tides

Besides the role as a mixing agent, tidal currents may also turn out to be an important transport mechanism for species, especially for passive drifters. Several species have, at an early stage in their life cycle, a diurnal migration pattern. Therefore a strong diurnal tidal component may lead to a substantial net drift: During a two week interval there will always be a period when maximum tidal currents occur near noon, and such matches the diurnal migration pattern to give a maximum displacement. Therefore, in some cases, the tidal current amplitude determines the maximum distance between spawning grounds and juvenile nursery areas. Also it has been suggested that in a horizontal temperature or salinity gradient some species migrate vertically to be able to move towards more favourable environmental conditions.