Lecture Notes in Oceanography

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http://www.es.flinders.edu.au/~mattom/IntroOc/index.html

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Contents

Wave measurements
Tide gauges
Remote sensors
Shear probes

Introduction: an opening lecture

For many years the Flinders University of South Australia has offered a first year topic Earth Sciences in two parts. The first semester topic, Earth Sciences 1A, covered the place of the Earth in the universe, aspects of geology, and an introduction into geophysics and hydrology. Meteorology and oceanography were covered in the second semester topic Earth Sciences 1B.

Beginning in 2000 the two topics are delivered as Earth Sciences 1, which continues as the first semester topic with identical content, and Marine Sciences 1 as the second semester topic. Marine Sciences 1 still contains extensive material on meteorology and physical oceanography but also contains an elementary introduction to aspects of marine biology.

These notes represent the topic content for physical oceanography. In addition, two introductory lectures place the atmospheric and oceanographic aspects of the topic in the context of the exact sciences; they are an abbreviated version of the first two lectures given at the beginning of the semester.

General aims and objectives

- To give students practical experience in general scientific methodology including laboratory and field-based experimentation and scientific report writing.
- To help students develop an understanding of the unifying principles and processes which are critical in understanding both the evolution and behaviour of the planet Earth, with a particular focus on aspects relating to the atmosphere and the ocean.
- To encourage critical thinking and assist in the development of both quantitative and qualitative problem solving skills.
- To assist in the development of communication and team work skills in a technical environment.

Specific syllabus objectives

- To provide an overview of the processes which determine the state of the atmosphere and the ocean and their dynamics.
- To describe our planetary environment and the governing cycles and processes which control its behaviour.
- To describe the processes and phenomena which directly affect the nature and behaviour of the "fluid" Earth, namely, the composition of the atmosphere and of seawater, the balance of forces which controls winds, ocean currents and waves in both media, and their role in climate.
- To describe the natural atmospheric and oceanic processes which impact on use of the Earth's resources including industrial, commercial and recreational use and conservation.
- To let students appreciate the importance of scientific understanding of physical processes in ocean and atmosphere and the forces behind them for environmental pollution and management problems.

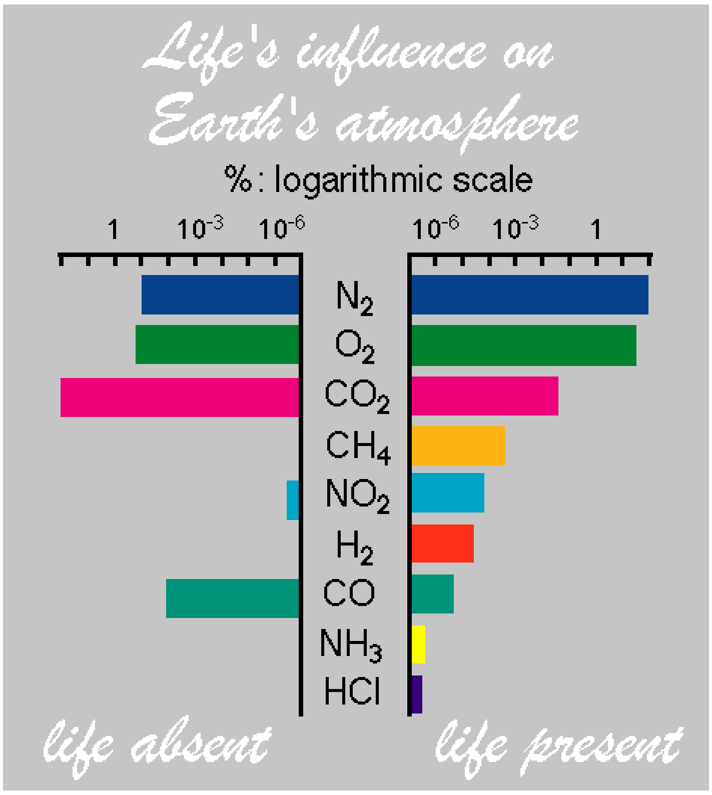
What will we learn today?

1. The environment Earth is shaped by the presence of life.
2. Understanding the environment means understanding the interaction between biosphere, geosphere, hydrosphere and atmosphere.

3. Earth sciences study the three non-living components of this interactive system.
4. Geosphere, hydrosphere and atmosphere are fluids in motion; their main difference is their viscosity.
5. As fluids in motion they exhibit some common features. Examples are convection, eddies, and energy transport through wave propagation.

The sciences of meteorology and oceanography study the results of these processes in the atmosphere and in the ocean.

The living and non-living components of the interactive system Earth are shaped through interaction - even the structure of its non-living components is determined by the presence of life. This is most pronounced for the atmosphere; its composition is determined by the presence of life.

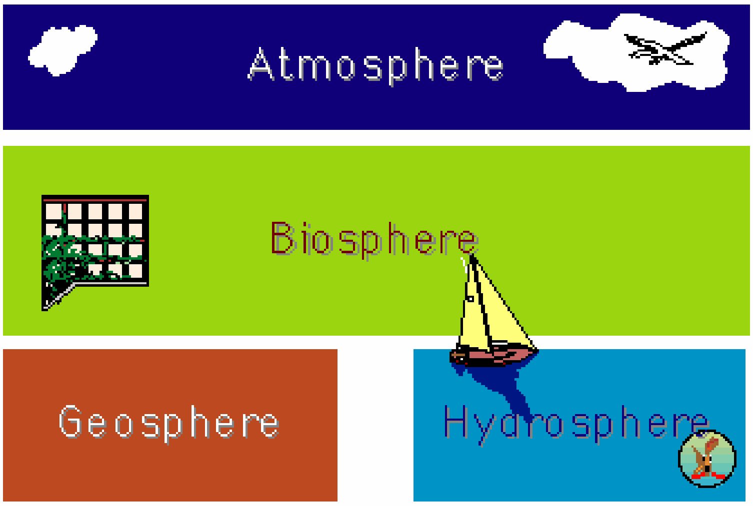


In the absence of life the most important gas in the atmosphere is carbon dioxyde (CO_2) which makes the atmosphere toxic for higher life forms. Life on Earth evolved through the activity of bacteria which reduced CO_2 levels to tolerable concentrations and produced oxygen (O_2) in the process.
Examples for this planetary state are Venus and Mars; both have an atmosphere with a CO_2 content of 95%.
The present make-up of Earth's atmosphere is maintained by continuous interplay between reduction of CO_2 and production of O_2 by plants and reduction of O_2 and production of CO_2 by animals and humans.
Without the presence of life the atmosphere would not maintain its composition. The earth's atmosphere is therefore said to be in "dynamic equilibrium", while in the absence of life atmospheres are said to be in "dead equilibrium".

In the 1960s the British chemist James Lovelock was engaged by the space agency NASA of the USA in its search for extra-terrestrial life, particularly on Mars and Venus. Noticing the large difference in atmospheric composition between Earth and the other two planets - and in particular the fact that the atmospheres of Venus and Mars are in a dead equilibrium while the atmosphere of Earth is in a dynamic equilibrium, which without life would immediately revert to the dead equilibrium - he and Lynn Margulis, a microbiologist from the USA, developed the **gaea** concept or **hypothesis**. They point out that the presence of life has far reaching consequences for the planet as a whole. The development of forest, for example, reduces the albedo (the reflectivity of the Earth's surface) considerably. As a result the Earth is several degrees warmer than it would be without the presence of life. The gaea hypothesis therefore states that the planet Earth is a living organism itself, and oceans, land, air and all lifeforms are different organs of a living body.

Whether one accepts the gaea hypothesis in its extreme formulation or not, it is beyond doubt that the gaea hypothesis is a scientific hypothesis and can withstand the many attempts to turn it into a "new age religion". The fact remains that Earth as it is today is determined in its physical state (the distribution of water and ice, the composition of the atmosphere, the weathering processes of rocks, and much more) by the presence of life.

Modern science has been extremely successful in explaining the Earth by dividing it into compartments which can be studied separately.



Modern science divides the Earth into living and non-living compartments. In reality Earth is an interactive system in which all elements are related and influence each other.
Not all cultures see the Earth that way; but the success of modern science shows that dividing the Earth into independent compartments can be one way of understanding many aspects of it.
This topic will follow the tradition of modern science; but students should keep in mind that the system *Earth* is more than the sum of independent parts.

It is not well equipped to approach Earth as an interrelated organism. This "western" way of analyzing and understanding the world is also reflected in the structure of western languages, which compose sentences through subject-object relationships wich always establish a clear master-servant hierarchy. A sentence such as: "The engineer improves the environment" says that there is an environment, of which the engineer is not a part; he is the master of the environment. Other cultures do not divide the world into compartments, and their languages describe the world in totally different ways. There are many examples of American Indian languages which do not know the concept of subject and object; and if the Australian Aborigines say: "We are the land, and the land is us." they express their view that dividing the world into compartments can make you loose sight of important interactions between the various "spheres."

Keep in mind that meteorology and oceanography are just two compartments of a system with many living and non-living interactive components and that studying processes through meteorology and oceanography is only one way of studying the world. Nevertheless, the success of western science in explaining how the physical world works should not be dismissed lightly, and we shall follow its methods.

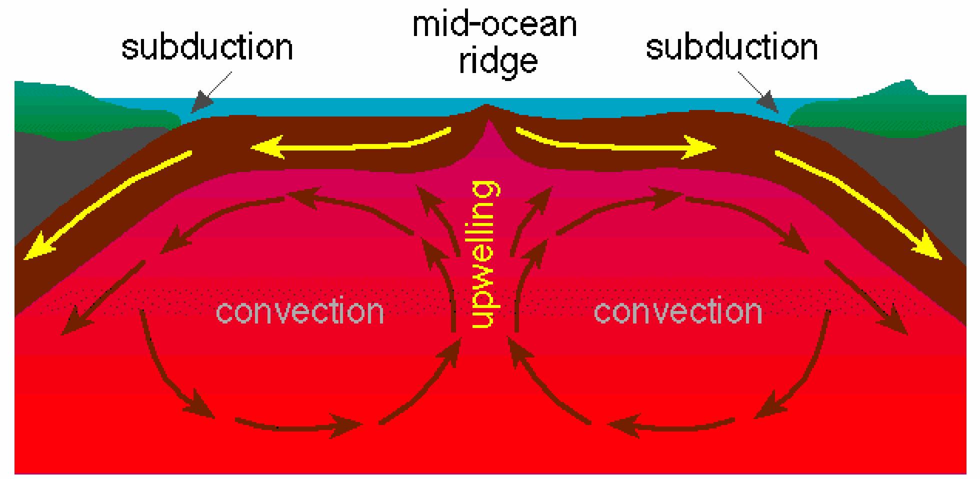
With this proviso, let us proceed to the study of physical processes in nature and look at three examples: convection, eddies and waves.

Convection

A fluid can be stratified, which means that its density can vary. For a fluid to be in a stable state, its density has to decrease from the bottom upwards to the top.

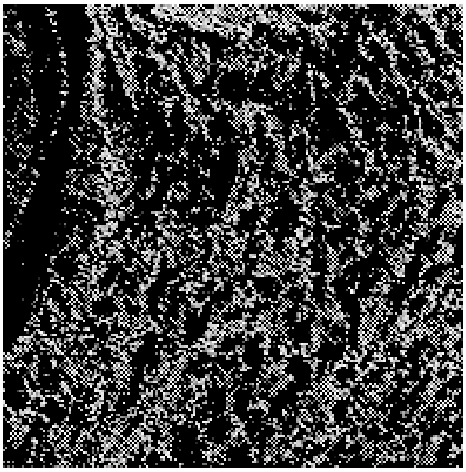
Convection occurs when this condition is not satisfied. Instability occurs when the density of the fluid is higher at the top than at the bottom. The lighter fluid then rises to the top, the denser fluid sinks to the bottom until stability is achieved. The resulting movement is called convection.

Convection represents a balance of forces between gravity and friction. A third force is required to establish the initial instability. The space and time scales of convection depend on the viscosity of the fluid. The examples given in the figures show convection in the "solid" earth, atmosphere and ocean.



Heating from the Earth's core drives convection in the upper mantle. This convection is extremely slow; the speed with which material in the Earth's crust spreads from the mid-ocean ridges is of the order of several cm per year. Nevertheless, it is evident that the same forces which drive convection in the atmosphere and in the ocean are present in the "solid" earth as well.

Convection in the atmosphere, observed through cloud formation and rain development in the equatorial Indian Ocean. In the atmosphere convection is driven by the ocean or the land surface, which receive heat from the sun during the day and in return heat the air from below. This lowers the air density near the ground and forces the air to rise. When a convection cell develops, the air rises in a region of about 10 kilometres. Cooling of the rising air causes condensation, so the vertical extent of the convection cell becomes visible as a tall cumulonimbus cloud. Sinking of air occurs around the cloud, so the outer region of the convection cell remains cloud free. As the air rises higher and higher (to 10 km height or more) the condensation turns into rain, which falls out as a heavy downpour from the centre of the convection cell. Convective rain accounts for most of the rainfall in the tropics. It is very intense when it falls but very patchy in space and not long lasting in time.



In the ocean convection occurs when the water at the sea surface is cooled. This causes an increase in water density at the surface and forces the water to sink. Because intense cooling is required to generate oceanic convection, the process occurs mainly in polar regions and is not as easily observed as atmospheric convection in the tropics. The photo shows a synthetic aperture radar image from the ERS-1 satellite over the Greenland Sea during February 1992. The image is about 6.4 km wide in both directions. Ice-covered water is grey/white, ice-free water is black. A noticeable feature is the patchiness of the ice. Ice-covered regions are about 500 m in diameter and surrounded by ice-free regions of similar size. This is interpreted as a result of convection: When a convection cell develops, water sinks with great speed. Water is drawn in to fill the void and accumulates ice floes over the regions of sinking water. Rising water motion occurs at the perimeter of the convection cells. This water is warmer than the cooled surface water and melts the ice where it comes to the surface. This produces a patchwork of ice-covered and ice-free regions.

Eddies

Eddies are the results of instabilities in fluid motion. They involve a somewhat more complicated balance of forces than what we intend to study here, but they are such common features that it is instructive to look at some examples and compare again the "solid" earth, the atmosphere and the ocean. The similarity of eddies in the atmosphere and in the ocean will be discussed in more detail in Lecture 1 later in this course. In this context it is worth noting that the "solid" earth undergoes very similar processes, although on much longer time scales.

In the atmosphere and in the ocean eddies can be generated from wind shear or current shear, ie when the fluid moves in the same direction but with different speed. The high viscosity of the "solid" earth often prevents eddy formation even when there is shear in the movement of the mantle or crust. Folding is observed instead.

A composite satellite image of cloud coverage over the earth. Several large eddies are clearly visible, among them a tropical cyclone (A) and a tropical depression (B) which may develop into a cyclone.

An example of a vortex street in water, produced in a laboratory experiment. The eddies develop as a current passes an obstacle. Similar eddies are observed in the ocean behind islands and in the atmosphere behind high mountains.

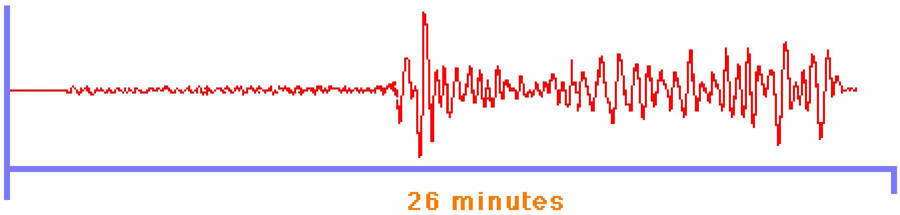
An example of strong bending of originally horizontal layers in the Canadian Rocky Mountains. If the forcing continues long enough the layers will eventually fold over and attain a shape resembling the water swirls in eddies.

Waves

Waves are a balance of forces where the forces vary periodically in strength and produce periodic fluid motion as a result. They are an efficient means to transport energy over large distances.

There will be ample opportunity later in this topic to study the interaction of forces in wave motion in detail.

At this point we use waves as another example which demonstrates again that the "solid" earth, the atmosphere and the ocean are three different types of fluid in motion.



Movement of the earth during the 1906 earthquake of San Francisco. The instrument shows wave motion produced by the earthquake. The waves traversed the earth's mantle and core to arrive at Göttingen, Germany, where this record was taken. Earthquakes also produce waves that travel along the earth's surface. These take much longer and did not arrive in Göttingen until after this record was obtained. When they arrived, the instrument went off-scale.

Cloud patterns in which clouds are organised in bands are not an unusual sight. The photo shows a cloud pattern produced by an internal wave. The air rises with the crest of the wave, cooling in the process, which produces condensation (cloud formation). It sinks in the troughs, causing warming and evaporation of the cloud moisture. The wave crests then become visible as cloud bands. The photo was taken near Mount Lawson (30 km west of Brisbane, Australia) and the wave was travelling towards Brisbane city, while it was a fine still day on the ground.

Surface waves are among the most elementary and most easily observed features of ocean dynamics, so not much comment is needed here. Later lectures in this topic will explore other types of waves that are not as easily appreciated as the ocean's surf.

There are many ways in which waves can be generated. The examples shown in the figures represent different force balances. What they have in common is that their properties can be understood and their behaviour predicted on the basis of the Laws of Physics.

What did we learn today?

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4. Geosphere, hydrosphere and atmosphere are fluids in motion; their main difference is their viscosity.
5. As fluids in motion they exhibit some common features: convection, eddies, energy transport through wave propagation (among others).

What will follow in this topic?

In today's world human activity - be it industrial, commercial or recreational - will shape our environment more than ever before. Active environmental management on a global scale has become a necessity.

Human activity cannot override the Laws of Nature. Active environmental management has to be based on these laws, it cannot succeed if it attempts to oppose them.

The sciences of meteorology and oceanography investigate and explain how the Laws of Physics determine processes in the atmosphere and oceans. They form the basis for any environmental management.

Responsible environmental management takes into account many factors, such as economical, social and historical considerations; but it cannot ignore the Laws of Physics.

Qualitative description and quantitative science

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The topic for today:
The concept of cycles and budgets

Meteorology and oceanography are physical sciences which aim to understand processes in the environment and describe, analyze and predict them in a quantitative manner.

A common way of expressing processes quantitatively is through the concept of cycles and budgets.

On time scales of geological history, all processes on earth are based on a constant reservoir of materials.

The forms in which the materials are present change constantly. In a state of equilibrium this change has to be cyclic.

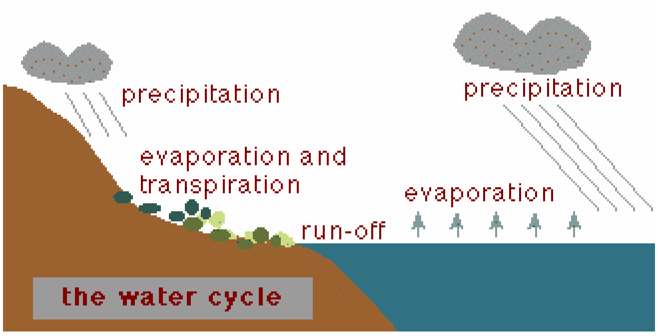
The concept of cycles expresses this principle of equilibrium in a qualitative manner. The concept of budgets makes it quantitative by giving rates of change between different states in the cycle.

This lecture discusses four examples.

The Water Cycle

The earth is the only planet in the solar system where liquid water is found on the surface. Water is the only substance which, under the ranges of pressure and temperature experienced on earth, is present in solid, liquid and gaseous phase. The water cycle is therefore of fundamental importance to many processes unique to earth. In comparison, the outer planets of our solar system (Saturn, Jupiter, Uranus, Neptune and Pluto) and their moons are too cold to contain water in any form other than ice, the inner planets (Mercury and Venus) are too hot to hold water in any form other than water vapour, and Mars is presently too cold but may have had liquid water on its surface at some point in its history. In the current stage of development of the solar system Earth is the only planet that contains water in all its phases.

Like many other cycles, the water cycle links processes acting in the living and non-living world: Precipitation and oceanic evaporation link ocean and atmosphere; evaporation from land and transpiration from vegetation link the atmosphere with the biosphere.



A sketch of the water cycle. Water cycles from the ocean to the atmosphere through evaporation, is transported in condensed form as clouds with the winds and returns to land and water as precipitation. The biosphere plays an important role in the water cycle. Evapo-transpiration from plants is the major component of the water cycle's pathway from the land to the atmosphere; direct evaporation from land is comparatively small. Evaporation from the sea, however, is quite significant. The link between land and ocean is represented by run-off from rivers. This closes the water cycle.

In the context of meteorology and oceanography the effect of the biosphere is quantitatively expressed as a single process, evapo-transpiration. The water cycle then describes a basic component of the combined system ocean-atmosphere.

Associated with every **cycle** is a **budget**. Cycles represent a *qualitative description* of processes, budgets turn them into *quantitative statements*. We distinguish between static budgets, which summarize how much of a particular material is available and how it is distributed between the different compartments, and dynamic budgets, which quantify how rapidly the material is moved between compartments. Cycles define the process; budgets allow answers to questions such as; "How is the water cycle affected if a given percentage of the existing bushland in Western Australia is cleared and replaced by wheat farming?"

The Water Budget

The distribution of water on earth (the static budget); this budget shows where the water is found:

region	volume (10^3 km^3)	% of total
oceans	1,350,000	94.12
groundwater	60,000	4.18
ice	24,000	1.67
lakes	230	0.016
soil moisture	82	0.006
atmosphere	14	0.001
rivers	1	-
Based on M. J. Lvovich: World water balance; in: Symposium on world water balance, UNESCO/IASH publication 93, Paris 1971.		

The static budget demonstrates the importance of the ice sheets to the global water cycle: Any change in the atmospheric or oceanic conditions that releases a significant part of the water that is presently stored in the ice, will produce a major shift in the water cycle. The atmosphere seems insignificant in comparison. However, the important role of the atmosphere becomes clear when the dynamic budget is considered.

The Water Flux Budget

The branches of the water cycle on earth (the dynamic budget); this budget shows how water moves between atmosphere and hydrosphere:

process	amount (m^3 per year)
precipitation on ocean	$3.24 \cdot 10^{14}$
evaporation from ocean	$-3.60 \cdot 10^{14}$
precipitation on land	$0.98 \cdot 10^{14}$
evaporation from land	$-0.62 \cdot 10^{14}$
net gain on land = river run-off	$0.36 \cdot 10^{14}$

The flux budget demonstrates that most of the water exchange between the compartments is between the ocean and atmosphere, so the atmosphere is an extremely dynamic element in the system despite of its small water content at any one time. The turnover of water between ocean and atmosphere over a few decades is equivalent to the total amount of water stored in the ice sheets.

The Salt Cycle

The salt cycle involves the ocean, the geosphere and to a very minor extent the atmosphere.

Minerals are leached from rocks through flowing groundwater and surface erosion. They enter the rivers and from there the ocean where they accumulate, making sea water salty. They are removed from the water and enter the sediment by chemical action.

The sediment is used to form new rock which brings the minerals back into the geosphere.

Salt gets into the atmosphere as spray from wind waves. This may be carried on to land, constituting a minute pathway from sea to land in the global salt cycle.

Because the salt cycle operates on such large time scales, establishing a static salt budget is of no relevance to oceanography.

Elements of the Salt Flux Budget

The salt cycle operates on such long scales that establishing a salt flux budget is not an important task for oceanography. The following table gives an idea of the time scales involved:

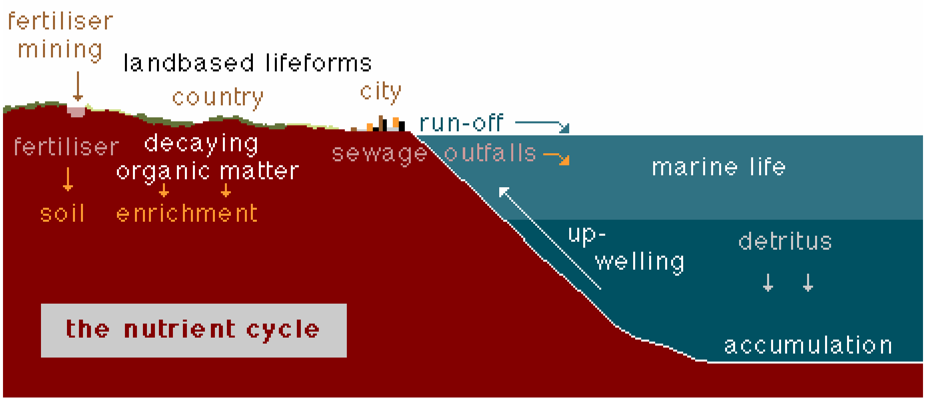
element	crustal abundance (%)	residence time (years)
some major constituents of sea salt:		
sodium (Na)	2.4	60,000,000
chlorine (Cl)	0.013	80,000,000
magnesium (Mg)	2.3	10,000,000
some trace constituents of sea salt:		
lead (Pb)	0.001	400
iron (Fe)	2.4	100
aluminium (Al)	6.0	100

The concept of salinity is the topic of lecture 3.

The Nutrient Cycle

Nutrients are essential for plant and animal life. They undergo a terrestrial and an oceanic cycle.

On land nutrients are taken up from the soil by plants and return to the soil by decomposition of dead organic matter. This is a closed cycle on a relatively short time scale, determined by the process of decomposition and life spans of plants, animals and humans. In developed human societies it is only broken by the uptake of nutrients by populations of large cities, which do not return the nutrients to the land but dispose of them in sewage systems. The resulting nutrient loss in agriculture is compensated by the importation of mineral fertiliser from the reservoir of minerals in the geosphere.



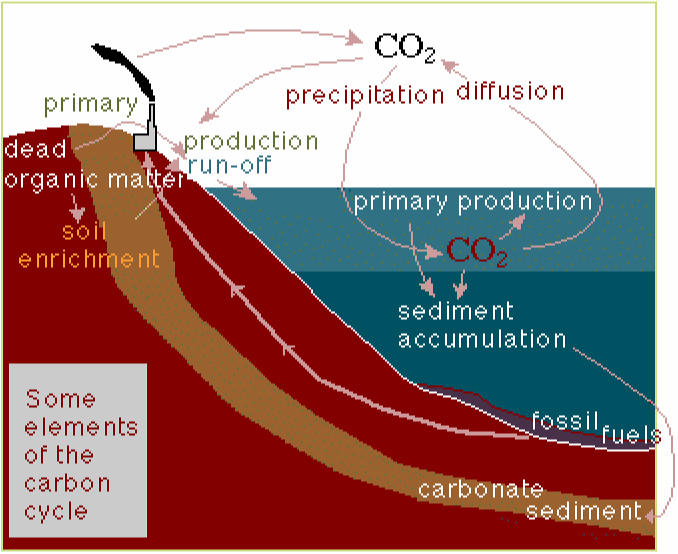
A sketch of the nutrient cycle. The natural cycle consists of the recycling of decaying organic matter into landbased lifeforms on land and nutrient supply for marine from upwelling in the ocean. The development of human civilisation introduces the additional elements of sewage disposal and fertiliser application.

This human influence introduces a link with a nutrient cycle of a much longer time scale, determined by the formation of mineral deposits. The situation is similar to the situation discussed with the carbon cycle below but does not have the same immediate consequences; the increase of nutrients available for the fast nutrient cycle on which life processes and agriculture depend is very gradual, and much of the mineral input is removed from the rapid nutrient cycle through the oceanic component.

In the ocean nutrient uptake by plants occurs in the surface layer reached by sunlight where photosynthesis takes place. Most nutrients are removed from the euphotic zone and transferred to the deeper ocean as dead organisms sink to the ocean floor, where they leave the rapid nutrient cycle. In the deeper layers organic matter is remineralized, i.e. nutrients are brought back into solution. Thus, the ocean cannot support highly productive ecosystems except where nutrients are returned to the euphotic zone from below in so called upwelling regions. The nutrient cycle is discussed in more detail in lecture 5, upwelling in lecture 6.

The Carbon Cycle

The carbon cycle operates naturally on two vastly different time scales. It involves the ocean, the atmosphere, the geosphere and the biosphere.



A sketch of the carbon cycle. A static budget would show that the largest amount of carbon is contained in the deepr layers of the geosphere (fossil fuels and carbonate sediment). The dynamic budget would show that it is stored there, while all day to day exchange of carbon between the compartments is active between the biosphere, the atmosphere and the upper ocean and continental layer. The uptake and burning of fossil fuels in power stations and cars establishes a link between the slow and fast cycles and leads to an increase of carbon dioxide in the atmosphere and in the ocean and a reduction in the geosphere.

On the **geological time scale** carbon is released into the atmosphere and ocean through the weathering of carbonate rocks such as limestones. It returns to this vast storage reservoir as new rocks are formed through sediment deposition.

On the much shorter **climate timescale** carbon is exchanged between the atmosphere, the ocean and living and dead organisms.

The carbon cycle includes both timescales, but for most practical purposes the carbon budget and the carbon flux budget usually exclude the geological timescale.

This separation between the timescales has been significantly disturbed through the burning of fossil fuel. This adds carbon dioxide to the atmosphere and increases its ability to retain heat energy received from the sun (the greenhouse effect). The following tables give some current estimates for the carbon budget and the carbon flux budget.

The carbon budget

region	amount (Gt carbon; 1 Gt = 10^{15} g)	
	before anthropogenic change	after anthropogenic change
land plants	610	550
soil and humus	1,500	no change
atmosphere	600	750 (+3.4 per annum)
uper ocean	1,000	1,020 (+0.4 per annum)
marine life	3	no change
dissolved organic carbon	700	no change
mid-depth and deep ocean	38,000	38,100 (+1.6 per annum)

The carbon flux budget
Balancing sub-budgets are identified by (a) - (d).

from	to	amount (Gt carbon per year; 1 Gt = 10^{15} g)	
		natural	anthropogenic
atmosphere	land plants	100 **(a)**	
	ocean	74 **(d)**	18
land plants	atmosphere	50 **(a)**	
	soil and humus	50 **(a)**	
soil and humus	atmosphere	50 **(a)**	
deforestation	atmosphere		about 1.9
fossil fuel	atmosphere		about 5.4
ocean sink	upper ocean		0.4
	mid-depth and deep ocean		1.6
rivers	ocean	0.8	
upper ocean	atmosphere	74 **(d)**	16
	marine life	about 40 **(b)**	
	mid-depth and deep ocean	90 **(c)**	5.6
marine life	upper ocean	about 30 **(b)**	
	mid-depth and deep ocean	4 **(b)**	
	dissolved organic carbon	6 **(b)**	
dissolved organic carbon	mid-depth and deep ocean	6 **(c)**	
mid-depth and deep ocean	upper ocean	100 **(c)**	
	sediment	0.13	

What did we learn today?

1. The state of the environment is determined by a balance of forces.

Defining several cycles, such as the water cycle, the salt cycle, the nutrient cycle and the carbon cycle, is a useful way of describing the equilibrium which results from the balance of forces.

2. The concept of cycles helps to understand the world; but to manage the environment and avoid mistakes the concept has to be turned into quantitative measurement.

Budgets and flux budgets turn the concept of cycles into quantitative statements.

Lecture 1

The place of physical oceanography in science; tools and prerequisites: projections, ocean topography

The atmosphere and the ocean are both fluids in turbulent motion and follow the same physical laws. The Flinders University of South Australia recognises this by presenting meteorology and oceanography in the topic **Marine Sciences 1** as a unit.

The following notes correspond roughly to the oceanography content of **Marine Sciences 1**.

A variety of textbooks offers coverage of physical oceanography at an introductory level. Most of these include a description of all aspects of marine sciences (i.e. including marine biology, geology and chemistry). Most are useful reference texts for physical oceanography.

Textbooks which cover only physical oceanography are usually much more detailed in their coverage than what is needed in an introductory course, but students with particular interest in physical oceanography aspects of earth sciences are encouraged to consult them as additional reference books.

The classification number for physical oceanography in the Dewy classification system is 551.46; the easiest way to find what the library has to offer in this field is to walk up to the shelves labelled 551.46 and browse through the books.

The place of physical oceanography in science

Physical oceanography occupies a unique place amongst all science disciplines because it has strong interactions with a large number of other sciences of very different characteristics. Universities usually follow one of two models in teaching physical oceanography. The first model emphasises the relationship between physical oceanography and other earth sciences disciplines:

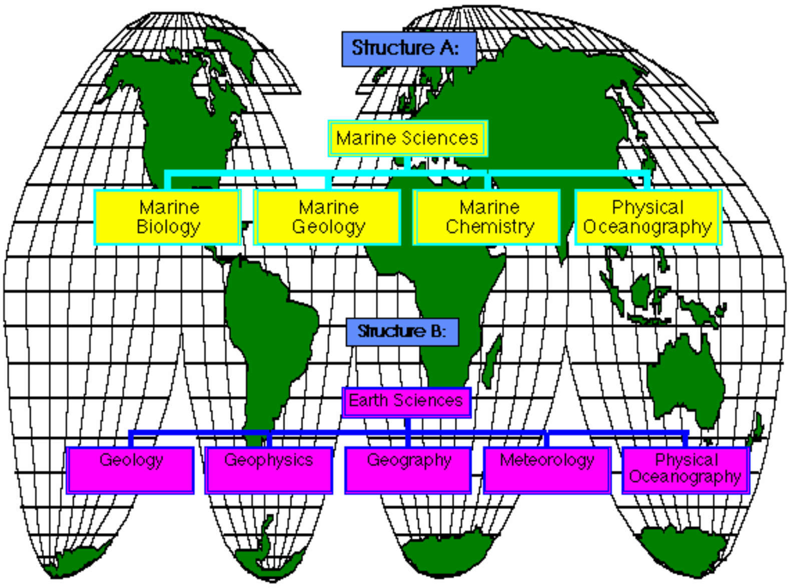


Figure. Structure of Earth Sciences and Marine Sciences (Structures A, B).

The common characteristics of the earth sciences are that they study components of the planet earth and try to understand how they work, i.e. how the laws of physics and chemistry act to shape the earth as it is observed today and as it was in the past. In contrast to other sciences (physics, chemistry, biology), earth sciences can rarely control the conditions of their experiments. Their task is to collect data from the field and interpret them to the best of their abilities.

The second model groups physical oceanography together with all other marine science disciplines. The common characteristics of the marine sciences are the use of special research tools to study the oceans such as research vessels, submersibles, moorings and drifters. Some of the marine sciences also rely on successes and progress in other marine science disciplines. For example, marine biology often needs information about the physical environment; physical oceanography would not have the detailed knowledge about the deep ocean circulation without progress in ocean technology.

Physical Oceanography itself can be subdivided into three major streams:

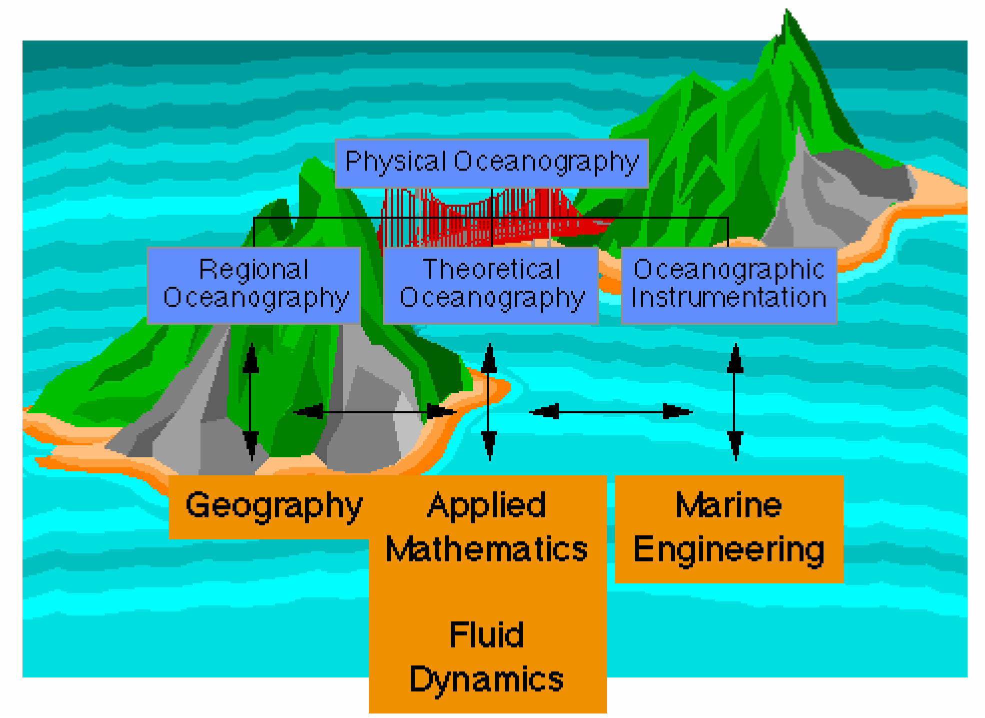


Figure. Structure of Physical Oceanography.

At Flinders University, all aspects of physical oceanography can be studied in a variety of ways depending on students' interests and strengths.

The study object of physical oceanography

The ocean is a fluid in turbulent motion, ie it is characterised by the presence of turbulent eddies with velocities often larger than the velocities of the mean flow. As the atmosphere is also a fluid in turbulent motion it can be expected that the two media, the study objects of physical oceanography and meteorology, show similar behaviour and are governed by the same balance of forces and that it is therefore advantageous to study both together. To demonstrate the similarity, Figure 1.1 and Figure 1.2 show examples of eddies in the atmosphere and in the ocean.

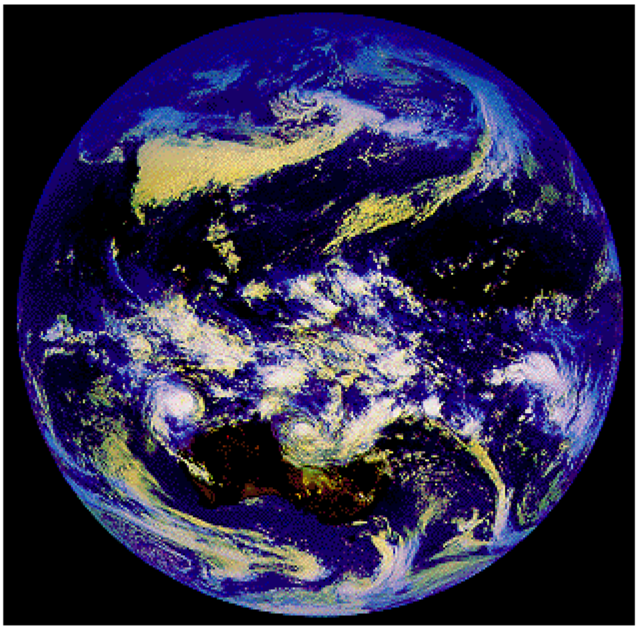


Figure 1.1. A view of the earth from a geostationary weather satellite. The turbulent character of the flow is evident from the cloud patterns, which act as tracers for the flow. It is seen that the flow is dominated by eddies, such as the eddy over central northern Australia.

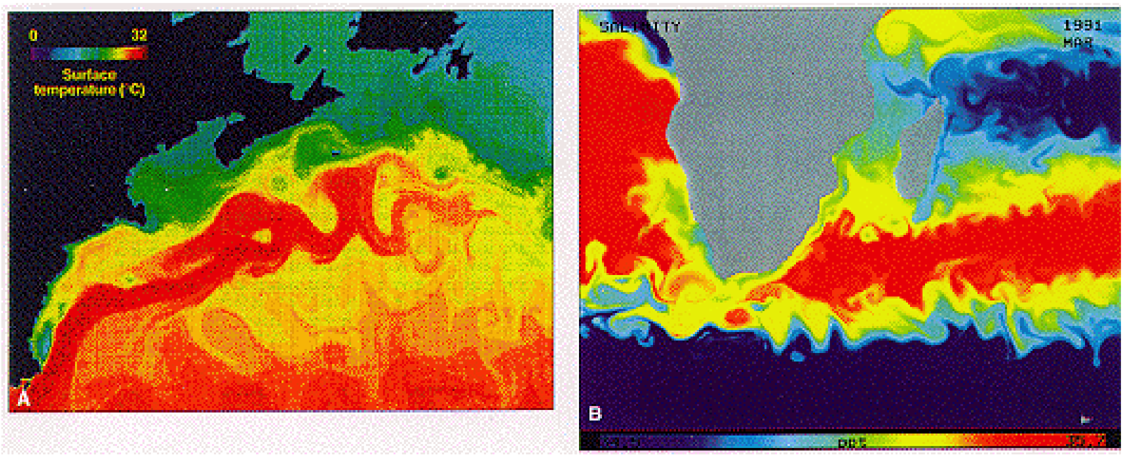


Figure 1.2. Two examples of eddies in a high resolution numerical model of the oceanic circulation. Left: The Gulf Stream as seen through sea surface temperature; right: eddies produced by the currents south of Africa as seen through sea surface salinity (dark is fresh, red is saline).

Note the difference in scale: Atmospheric eddies are typically some 2000 km in diameter, while oceanic eddy diameter are typically 200 km. A movie sequence of images taken from the weather satellite or from the ocean model would show that atmospheric time scales are also different: At any given location, atmospheric eddies pass at a rate of one eddy every 5 - 7 days (experienced as the passage of fronts), while oceanic eddy movement is such that the passage of an eddy takes 50 - 70 days.

The aim of oceanography is an understanding of the oceanic circulation and the distribution of heat in the ocean, how the ocean interacts with the atmosphere, and what role the ocean plays in maintaining our climate.

Tools and prerequisites for physical oceanography

Projections

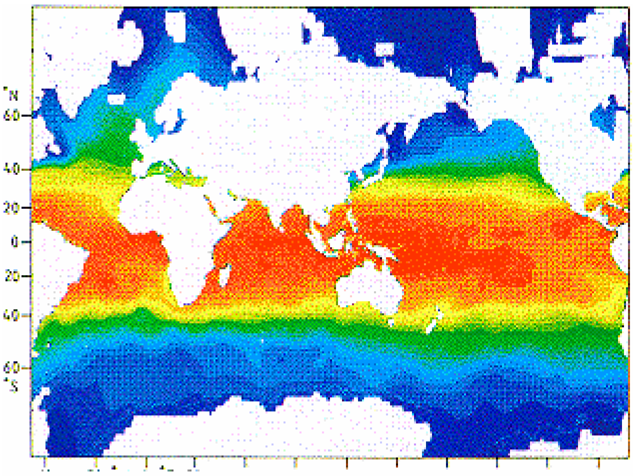
An important tool in oceanography (as in all other earth sciences) is the atlas. People are used to looking up items of interest in an atlas, but few realise the importance of the correct choice of projections used for the maps.

A projection widely used in physical oceanography is the Mercator projection. It was developed in the 16th century at a time of colonial exploration and expanded sea travel. Columbus had discovered America and Magellan's ships had circled the globe. One problem faced by these mariners was the uncertainty involved in navigation away from the coast. In the 16th century a navigator had to sail between two points along a rhumb line (a line of constant compass bearing) because it was not practical to do otherwise. Mercator developed a projection that showed the earth's surface in such a way that a straight line on the resulting plane anywhere and in any direction was a rhumb line. Thus a mariner with a knowledge of the starting position could draw a straight line to the destination and read off the correct bearing.

As a result, Mercator's projection has become the standard projection for navigation. It is, however, not an equidistant or equal area projection and therefore unsuitable for mapping over large areas. It is a conformal map, i.e. small circles of equal area on the earth are represented as circles on the map but increase in size towards the poles. The poles cannot be shown in a Mercator projection, since distances near the poles grow to infinity. In principle, representation of a curved surface on a plane always involves some "stretching" or "shrinking" resulting in distortions, or some "tearing" resulting in interruption of the surface. No projection can satisfy all three desirable properties, i.e.

- Equidistance - correct representations of distances
- Conformality (Orthomorphism) - correct representation of shapes
- Equivalency - correct representation of areas

The three criteria are basic but mutually exclusive. All other properties are of a secondary nature. Most projections with the property of fidelity of area achieve area conservation through the use of a curved longitude grid and therefore require a grid drawn over the map surface to enable determination of geographical location coordinates (Figure 1.3). The Gall/Peters projection, which was developed by Gall in 1855 and rediscovered again independently by Peters in the 1970s, combines fidelity of area with a rectangular latitude/longitude grid. It is ideal for mapping large ocean regions.



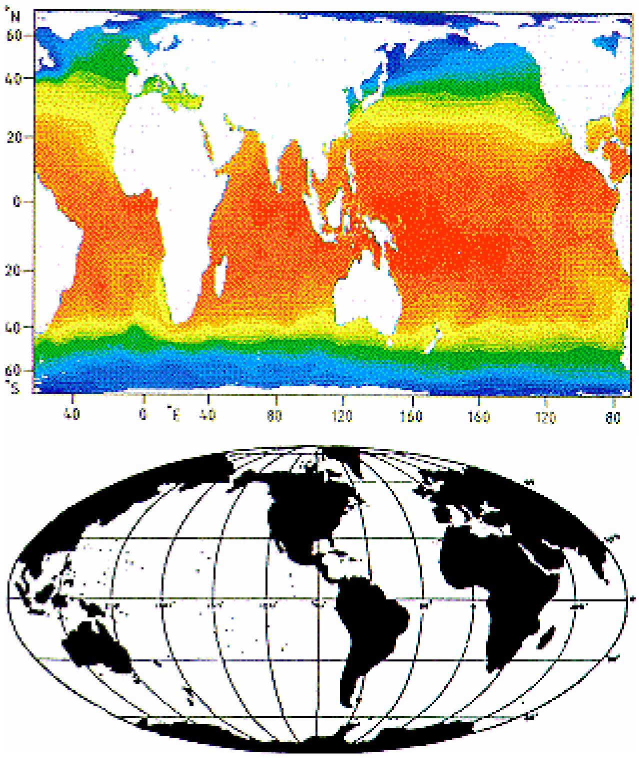


Figure 1.3. A comparison of three projections. Left: Mercator projection, middle: Gall/Peters (equal area) projection, right: an example of an equal area projection with curved coordinate grid (Mollweide projection). The coloured maps show sea surface temperature; they give *identical* information - differences of appearance are exclusively the result of different projections.
Compare the maps by locating the equator, noting the distance variation between successive latitudes, and trying to determine the longitude of your home town or university.

Topographic features of the oceans

The surface of the earth varies in height from 8848m (Mount Everest) to a depth of 11022 m (Vitiaz deep in the Mariana Trench, western North Pacific Ocean). On a geological time scale the position of the coastline depends on the amount of water available, which is mainly determined by the amount of ice and snow bound in Antarctica and in the Arctic Ocean, and to some degree by the temperature of the water in the ocean (water expands when warmed, so sea level rises during periods of warm climate). A characteristic of the land/water distribution of today, which has important ramifications for the climate, is that the area covered by water increases continuously from 70°N to 60°S:

Water coverage of the earth: northern hemisphere 61%, southern hemisphere 81%, global average 71%.

"land hemisphere" 53% (pole off the Loire river in France), "water hemisphere" 89% (pole near New Zealand).

The present distribution of land and sea and the various depth levels is shown in the so-called hypsographic curve (Figure 1.4).

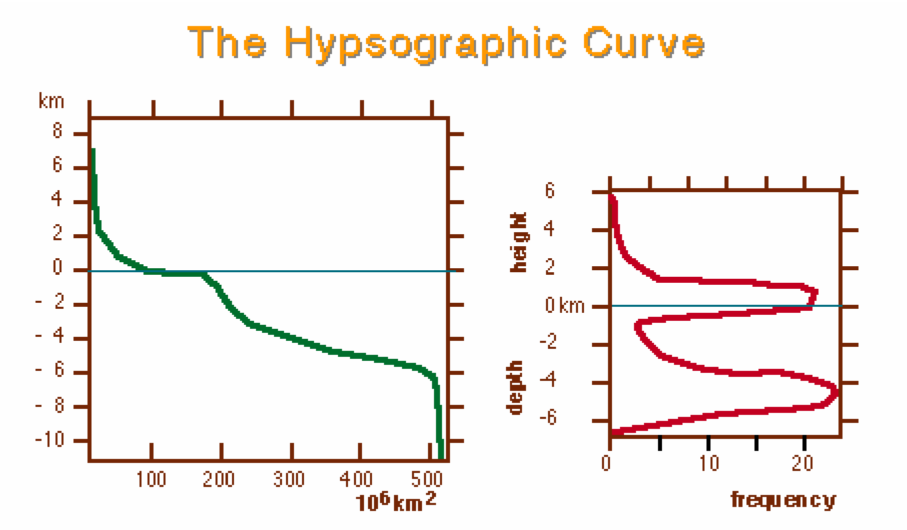


Figure 1.4. The hypsographic curve for the surface of the earth (left) and the corresponding frequency distribution of elevation (right). Note the distinctly bi-modal distribution in the elevation distribution. It is apparent that most of the earth's surface is covered by deep basins and low lying land.

Average elevation -2440 m; mean land elevation +840 m, average ocean floor level -3795 m

The major oceans are structured into continental margins, mid- ocean ridges and deep sea basins (Figure 1.5.).

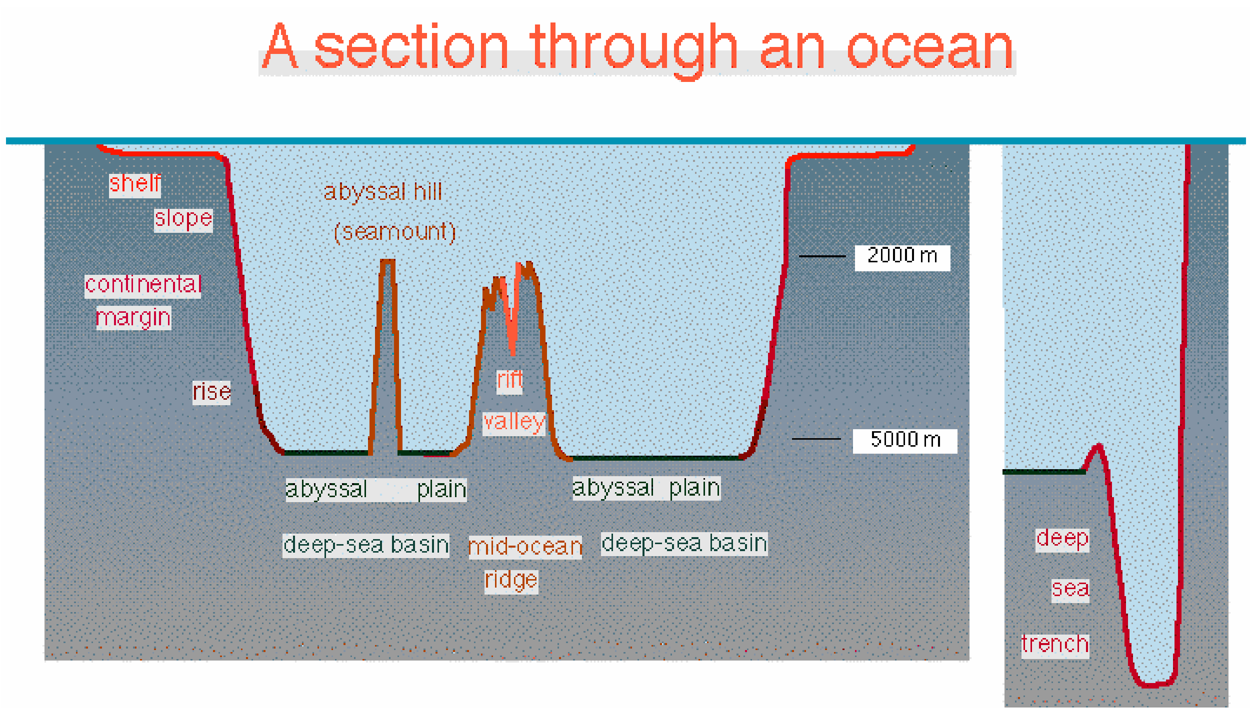


Figure 1.5. A cross-section through a typical ocean basin.

Each structural feature occupies about one third of the ocean floor.

topographic feature	width	depth	characteristics
Continental margins:			
Shelf	up to 300 km wide	150-200 m deep	
Slope	20 - 100 km wide	from 200 to 2000 m deep	Often furrowed by canyons. Slopes 1 in 40.
Rise	up to 300 km wide	from 2000 to 5000 m deep	Slopes 1 in 700 to 1 in 1000
Trench		600 to 11,000 m deep	There are 26 trenches in the world ocean: • 3 in the Atlantic Ocean • 1 in the Indian Ocean • 22 in the Pacific Ocean
Deep sea basins		about 5000 m deep	
Abyssal Plains			extremely flat, sediment-filled
Abyssal Hills			Rise from the plains up to 1000 m
Mid-ocean ridge: Interconnected mountain system	up to 400 km wide		Rises to 3000 - 1000 m
Central rift valley	20 - 50 km wide		cuts 1000 - 3000 m deep into the ridge system

Scales of graphs

As noted the average depth of the oceans is just under 4 km. This is quite deep. Or is it? If you use a divider and a very sharp pencil to draw a circle of 15 cm radius and take it to represent the earth then the pencil line would be thick enough to represent the crust of the Earth under continents (30 km) but much too thick for the oceanic crust (10 km). Irregularities in the line would be more than enough to represent the variation in the relief of the solid earth. So the ocean is really only a thin film of fluid - if the earth was a basketball, one would notice that much if its surface was damp.

There is no way of showing the oceans on a scale that preserves the aspect ratio of horizontal vs. vertical length. So how are we going to map ocean properties such as temperature, salinity or currents that vary considerably with depth? Compared to the ocean's vertical extent, horizontal distances are so large that the only way to produce a meaningful representation of the data is to distort the scales. A given distance on a diagram will thus represent several hundred times as much in real distance in the horizontal as in the vertical. A typical ratio is 500:1. This should be kept in mind when looking at oceanographic sections.

Lecture 2

Objects of study in Physical Oceanography

Physical oceanography studies all forms of motion in the ocean. It relates observations of motion to physical laws (such as Newton's Law that if a force **F** acts on a body of mass *m*, it undergoes an acceleration **a** such that **F** = *m* **a**).

(Note: For the mathematically inclined, **bold** characters in this and all following lectures indicate vectors, characters in *italics* stand for scalars. If you do not know what that means, don't despair; it is more a matter of using the correct notation than a matter of practical importance for this topic.)

The geographical and atmospheric framework

The prevailing wind system is the major driving force for ocean currents. Figure 2.1 shows that in the open ocean winds are nearly zonal (blowing east-west). The Trade Winds are easterly winds in the tropics and subtropics (between 30°N and 30°S). They are regions of extremely uniform wind conditions, where the wind blows steadily from the same direction with moderate strength throughout the year. Their strength increases slightly in winter. The Trade Winds of the two hemispheres are separated by the Doldrums, a region of weak and variable winds near 5°N.

Between 30° and 65° are the Westerlies. They are stronger in winter than in summer and are regions of frequent storms. Poleward of 65° the wind direction reverses again, and the Polar Easterlies blow from east to west.

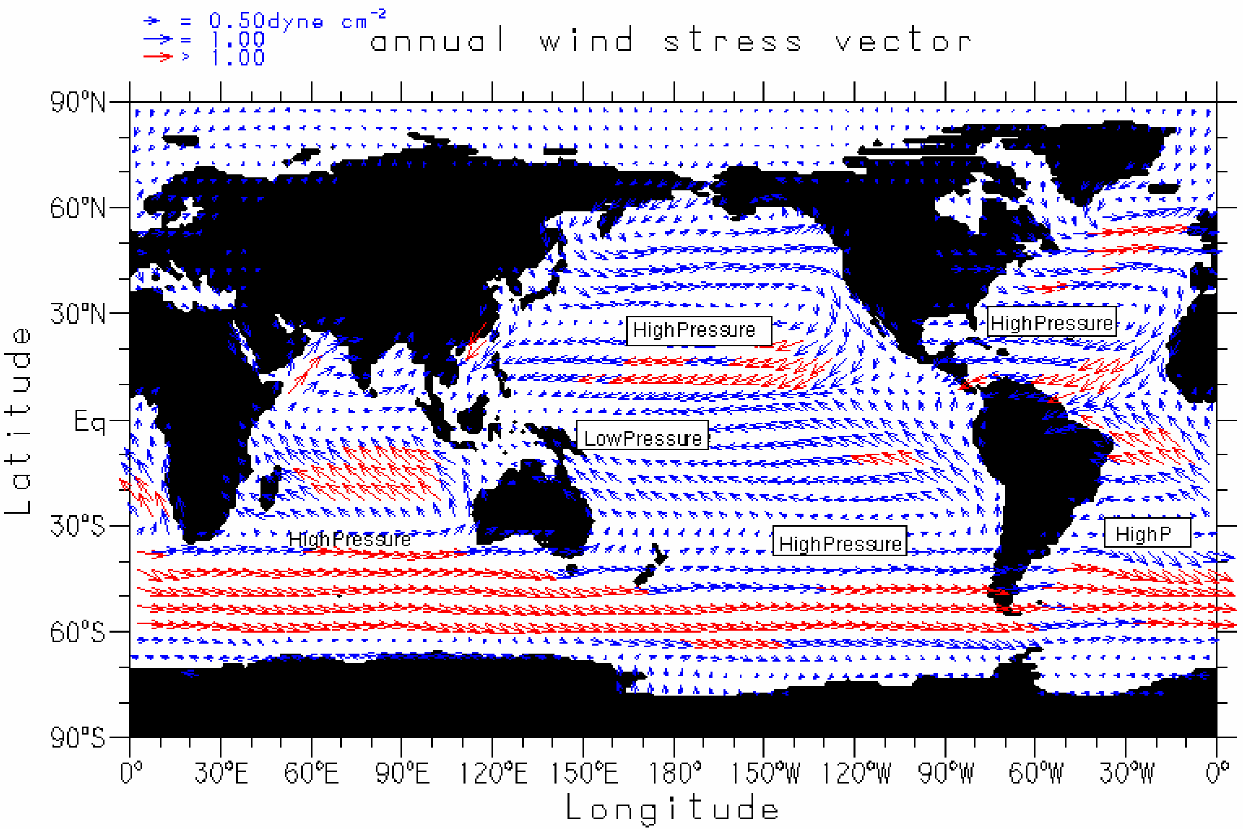


Figure 2.1. Annual mean surface wind stress over the oceans. From the course notes of General Oceanography by Don Reed, reproduced by permission.

Deviations from zonal wind direction occur near continents. This is particularly noticeable along the east coast of the oceans in the tropics and subtropics where the winds blow parallel to the coast towards the equator.

The present configuration of the distribution between land and water determines the ocean's response to the winds. It defines the major subdivisions of the world ocean, the Pacific, Indian and Atlantic Oceans. Their southern region around Antarctica is also known as the Southern Ocean.

Figure 2.2 is a map of surface currents.

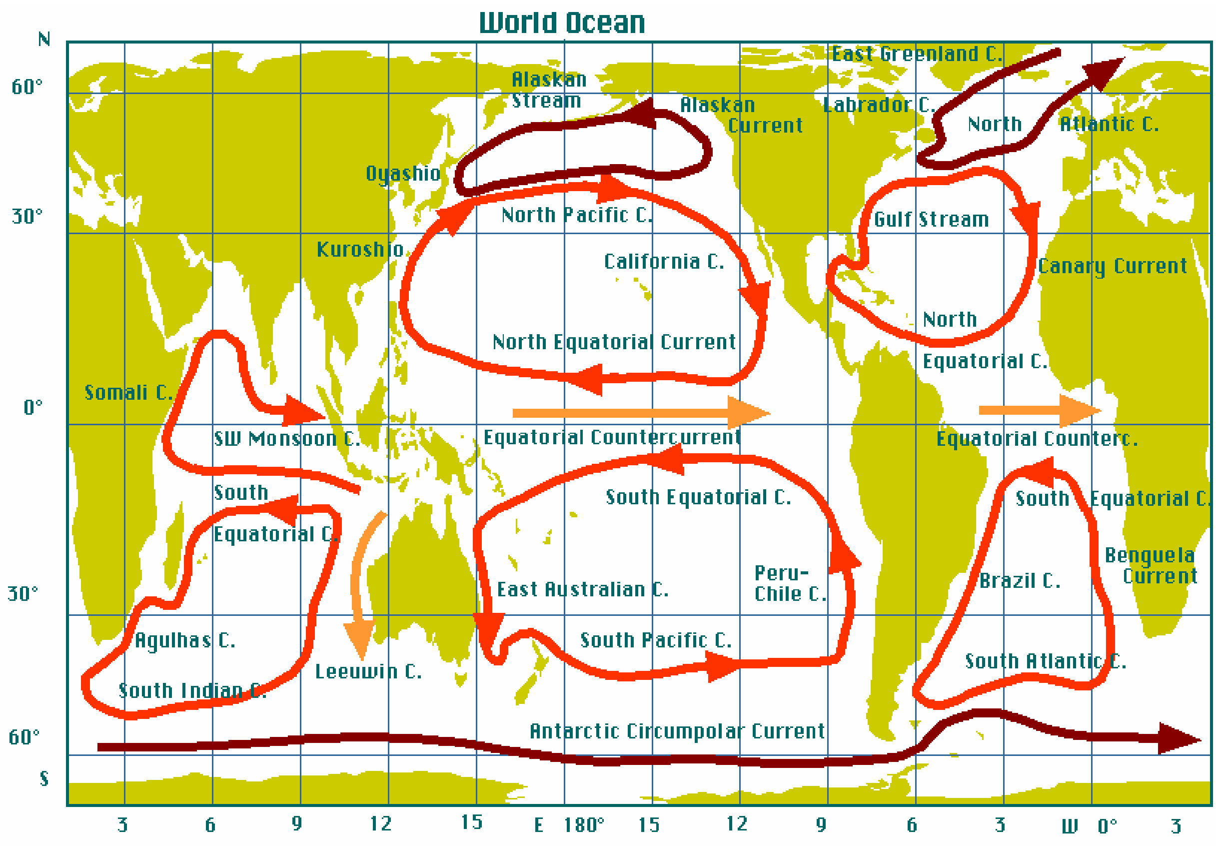


Figure 2.2. Surface currents of the world ocean. Currents in the Indian Ocean are shown for September/October (Southwest Monsoon season). The subtropical gyres are shown in orange, the subpolar gyres and the equivalent current in the southern hemisphere in dark brown. Tropical currents are shown in gold. Note that the figure does not indicate the strength of the current; western boundary currents are much stronger than all other currents.

The combined action of the Trade Winds and the Westerlies produces large gyres, with clockwise rotation in the northern hemisphere, anti-clockwise rotation in the southern hemisphere, known as the subtropical gyres. A subpolar gyre is produced in the north Pacific Ocean by the combined action of the Westerlies and the Polar Easterlies; it consists of the Oyashio, North Pacific Current and Alaskan Current. An indication of a subpolar gyre is also seen in the north Atlantic Ocean (anti-clockwise rotation in the current system that includes the North Atlantic, East Greenland and Labrador Currents). The subpolar region of the southern hemisphere does not have land barriers and therefore is dominated by the Antarctic Circumpolar Current.

> Note: The convention for indicating the direction of ocean currents differs from the convention used for wind directions. A "**westerly**" wind is a wind which blows from the west and goes to the east; a "**westward**" current is a current which comes from the east and flows towards west. This can cause confusion to people who rarely, if ever, go to sea; but it is easily understood and remembered when related to practical experience with winds and ocean currents. On land, it is important to know from where the wind blows: any windbreak must be erected in this direction. Where the wind goes is of no consequence. At sea, the important information is where the current goes: a ship exposed to current drift has to stay well clear from obstacles downstream. Where the water comes from is irrelevant.
> (M. Tomczak and J. S. Godfrey: Regional Oceanography: an Introduction. Pergamon, New York (1994), 422 pp.)
>
> So remember: **westerly** means from west, **westward** means towards west.

A feature to note is that as a general rule currents along the western coasts of ocean basins are much narrower and stronger than currents in the remainder of the ocean. Typical current velocities at the surface in the open ocean are 0.2 - 0.5 m s^{-1} (about 1 km h^{-1}). In western boundary currents velocities are closer to 2 m s^{-1} (about 7 km h^{-1}). These differences in current strength do not come out in most surface current maps.

The Indian Ocean is dominated by seasonal wind reversal (the Monsoons) and a corresponding reversal of surface currents. Figure 2.2 shows the situation during the Southwest Monsoon season when the Equatorial Countercurrent is suppressed and the circulation in the northern Indian Ocean differs significantly from that of the other ocean basins.

Just as in the atmosphere, where wind systems are linked to atmospheric pressure patterns, ocean currents are linked to pressure patterns in the ocean. Pressure at any depth in the ocean is determined by the weight of the water above, which is determined by the density of the water, which in turn depends on the water's temperature and salinity. It follows that ocean currents can be determined by measuring temperature and salinity, a task infinitely easier than direct current measurement. It is therefore appropriate to turn to a discussion of the basic properties of seawater before proceeding with a discussion of the oceanic circulation and the physical laws that govern it. This is the topic of the next lecture.

Lecture 3
Properties of seawater

Sea water is a mixture of 96.5% pure water and 3.5% other material, such as salts, dissolved gases, organic substances, and undissolved particles. Its physical properties are mainly determined by the 96.5% pure water. The physical properties of pure water will therefore be discussed first.

Pure water, when compared with fluids of similar composition, displays most uncommon properties. This is the result of the particular structure of the water molecule H_2O: The hydrogen atoms carry one positive charge, the oxygen atom two negative charges, but the atom arrangement in the water molecule is such that the charges are not neutralized (See Figure 3.1; the charges would be neutralized if the angle were 180° rather than 105°).

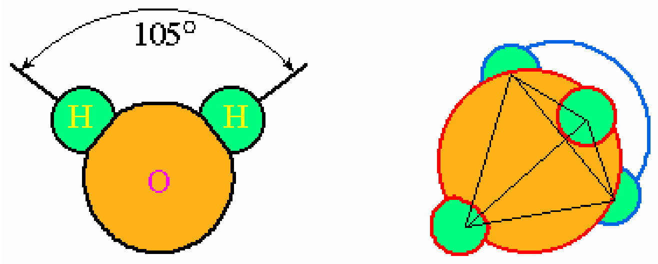


Fig.3.1. Left diagram: Arrangement of the oxygen atom (O) and the two hydrogen atoms (H) in the water molecule. The angle between the positively charged hydrogen atoms is 105°, which is very close to the angles in a regular tetrahedron (109° 28'). Right diagram: Interaction of two water molecules in the tetrahedral arrangement of the *hydrogen bond*. The hydrogen atoms of the blue water molecule attach to the red water molecule in such a way that the four hydrogen atoms form a tetrahedron.

The major consequences of the molecular structure of pure water are:

1. The water molecule is an electric dipole, forming aggregations of molecules (polymers), of on average 6 molecules at 20°C. Therefore, water reacts slower to changes than individual molecules; for example the boiling point is shifted from -80°C to 100°C, the freezing point from -110°C to 0°C.
2. Water has an unusually strong disassociative power, i.e. it splits dissolved material into electrically charged ions (Figure 3.2). As a consequence, dissolved material greatly increases the electrical conductivity of water. The conductivity of pure water is relatively low, but that of sea water is midway between pure water and copper. At 20°C, the resistance of sea water of 3.5% salt content over 1.3 km roughly equals that of pure water over 1 mm.

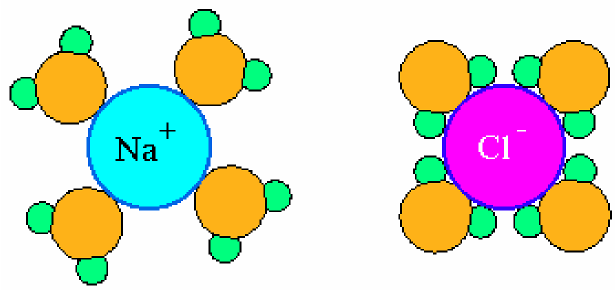


Fig. 3.2. Sodium chloride Na+Cl- is the main component of salinity in the ocean. The diagram shows how in hydrated form it causes water molecules to attach themselves with the positive hydrogen charges to the chloride and with the negative oxygen charge to the natrium.

3. The angle 105° is close to the angle of a tetrahedron, i.e. a structure with four arms emanating from a centre at equal angles (109° 28´). As a result, oxygen atoms in water try to have four hydrogen atoms attached to them in a tetrahedral arrangement (Figure 3.1). This is called a "hydrogen bond", in contrast to the (ionic) molecular bond and covalent bonding. Hydrogen bonds need a bonding energy 10 to 100 times smaller than molecular bonds, so water is very flexible in its reaction to changing chemical conditions.
4. Tetrahedrons are of a more wide-meshed nature than the molecular closest packing arrangement. They form aggregates of single, two, four and eight molecules. At high temperatures the one and two molecule aggregates dominate; as the temperature falls the larger clusters begin to dominate (Figure 3.3). The larger clusters occupy less space than the same number of molecules in smaller clusters. As a result, the density of water shows a maximum at 4°C.

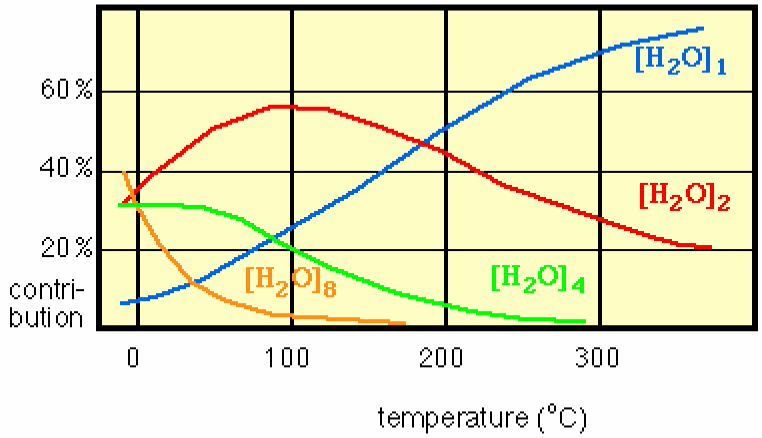


Fig. 3.3. The relative contributions of the different molecular aggregates of water as functions of temperature. The last index indicates the number of molecules in the aggregate.

Physical properties of most substances show uniform variation with temperature. In contrast, most physical properties of pure water show a minimum at some intermediate temperature. Sound velocity shows a maximum at 74°C (Table 3.1).

A list of some minimum temperatures	
The physical property is given first, followed by the temperature in °C at which the minimum occurs.	
oxygen solubility	80
specific volume	4
specific heat	34
hydrogen solubility	37
compressibility	44
speed of light	-1
speed of sound (maximum)	74

When freezing, all water molecules form tetrahedrons. This leads to a sudden expansion in volume, ie a decrease in density. The solid phase of water is therefore lighter than the liquid phase, which is a rare property. Some important consequences are:

1. Ice floats. This is important for life in freswater lakes, since the ice acts as an insulator against further heat loss, preventing the water to freeze from the surface to the bottom.

2. Density shows a rapid decrease as the freezing point is approached. The resulting expansion during freezing is a major cause for the weathering of rocks.
3. The freezing point decreases under pressure. As a consequence, melting occurs at the base of glaciers, which facilitates glacier flow.
4. Hydrogen bonds give way under pressure, i.e. ice under pressure becomes plastic. As a consequence, the inland ice of the Antarctic and the Arctic flows, shedding icebergs at the outer rims. Without this process all water would eventually end up as ice in the polar regions.

The Concept of Salinity

As mentioned before, sea water contains 3.5% salts, dissolved gasses, organic substances and undissolved particulate matter. The presence of salts influences most physical properties of sea water (density, compressibility, freezing point, temperature of the density maximum) to some degree but does not determine them. Some properties (viscosity, light absorbtion) are not significantly affected by salinity. (Particle and dissolved matter do affect light absorption in sea water and this influence is used in most optical applications.) Two properties which are determined by the amount of salt in the sea are conductivity and osmotic pressure.

Ideally, salinity should be the sum of all dissolved salts in grams per kilogram of sea water. In practice, this is difficult to measure. The observation that - no matter how much salt is in the sea - the various components contribute in a fixed ratio, helps overcome the difficulty. It allows determination of salt content through the measurement of a substitution quantity and calculation of the total of all material making up the salinity from that measurement.

Determination of salinity could thus be made through its most important component, chloride. Chloride content was defined in 1902 as the total amount in grams of chlorine ions contained in one kilogram of sea water if all the halogens are replaced by chlorides. The definition reflects the chemical titration process for the determination of chloride content and is still of importance when dealing with historical data.

Salinity was defined in 1902 as the total amount in grams of dissolved substances contained in one kilogram of sea water if all carbonates are converted into oxides, all bromides and iodides into chlorides, and all organic substances oxidized. The relationship between salinity and chloride was determined through a series of fundamental laboratory measurements based on sea water samples from all regions of the world ocean and was given as

$S\ (^{o}/_{oo}) = 0.03 + 1.805\ Cl\ (^{o}/_{oo})$ (1902)

The symbol $^{o}/_{oo}$ stands for "parts per thousand" or "per mil"; a salt content of 3.5% is equivalent to 35 $^{o}/_{oo}$, or 35 grams of salt per kilogram of sea water.

The fact that the equation of 1902 gives a salinity of 0.03 $^{o}/_{oo}$ for zero chlorinity is a cause for concern. It indicates a problem in the water samples used for the laboratory measurements. The United Nations Scientific, Education and Cultural Organization (UNESCO) decided to repeat the base determination of the relation between chlorinity and salinity and introduced a new definition, known as **absolute salinity**,

$S\ (^{o}/_{oo}) = 1.80655\ Cl\ (^{o}/_{oo})$ (1969)

The definitions of 1902 and 1969 give identical results at a salinity of 35 $^{o}/_{oo}$ and do not differ significantly for most applications.

The definition of salinity was reviewed again when techniques to determine salinity from measurements of conductivity, temperature and pressure were developed. Since 1978, the "Practical Salinity Scale" defines salinity in terms of a conductivity ratio:

" The **practical salinity**, symbol S, of a sample of sea water, is defined in terms of the ratio K of the electrical conductivity of a sea water sample of 15°C and the pressure of one standard atmosphere, to that of a potassium chloride (KCl) solution, in which the mass fraction of KCl is 0.0324356, at the same temperature and pressure. The K value exactly equal to one corresponds, by definition, to a practical salinity equal to 35." The corresponding formula is:

$$S = 0.0080 - 0.1692\ K^{1/2} + 25.3853\ K + 14.0941\ K^{3/2} - 7.0261\ K^2 + 2.7081\ K^{5/2}$$

Note that in this definition, salinity is a ratio and ($^{o}/_{oo}$) is therefore no longer used, but an old value of $35^{o}/_{oo}$ corresponds to a value of 35 in the practical salinity. Some oceanographers cannot get used to numbers without units for salinity and write "35 psu", where psu is meant to stand for "practical salinity unit". As the practical salinity is a ratio and therefore does not have units, the unit "psu" is rather meaningless and strongly discouraged. Again, minute differences occur between the old definitions and the new Practical Salinity Scale, but they are usually negligible.

Electrical Conductivity

The conductivity of sea water depends on the number of dissolved ions per volume (i.e. salinity) and the mobility of the ions (ie temperature and pressure). Its units are mS/cm (milli-Siemens per centimetre). Conductivity increases by the same amount with a salinity increase of 0.01, a temperature increase of 0.01°C, and a depth (ie pressure) increase of 20 m. In most practical oceanographic applications the change of conductivity is dominated by temperature.

Density

Density is one of the most important parameters in the study of the oceans' dynamics. Small horizontal density differences (caused for example by differences in surface heating) can produce very strong currents. The determination of density has therefore been one of the most important tasks in oceanography. The symbol for density is the Greek letter ρ (rho).

The density of sea water depends on temperature *T*, salinity *S* and pressure *p*. This dependence is known as the **Equation of State of Sea Water.**

The equation of state for an ideal gas was is given by

$$p = \rho R T$$

where *R* is the gas constant. Seawater is not an ideal gas, but over small temperature ranges it comes very close to one. The exact equation for the entire range of temperatures, salinities and pressures encountered in the ocean

$$\rho = \rho(T,S,p)$$

(where *S* is salinity) is the result of many careful laboratory determinations. The first fundamental determinations to establish the equation were made in 1902 by Knudsen and Ekman. Their equation expressed ρ in g cm^{-3}. New fundamental determinations, based on data over a larger pressure and salinity range, resulted in a new density equation, known as the "International Equation of State (1980) ". This equation uses temperature in °C, salinity from the Practical Salinity Scale and pressure in dbar (1 dbar = 10,000 pascal = 10,000 N m^{-2}) and gives density in kg m^{-3}. Thus, a density of 1.025 g cm^{-3} in the old formula corresponds to a density of 1025 kg m^{-3} in the International Equation of State.

Density increases with an increase in salinity and a decrease in temperature, except at temperatures below the density maximum (Figure 3.4). Oceanic density is usually close to 1025 kg m^{-3} (In freshwater it is close to 1000 kg m^{-3}). Oceanographers usually use the symbol σ_t (the Greek letter sigma with a subscript t) for density, which they pronounce "sigma-t". This quantity is defined as $\sigma_t = \rho - 1000$ and does not usually carry units (it should carry the same units as ρ).

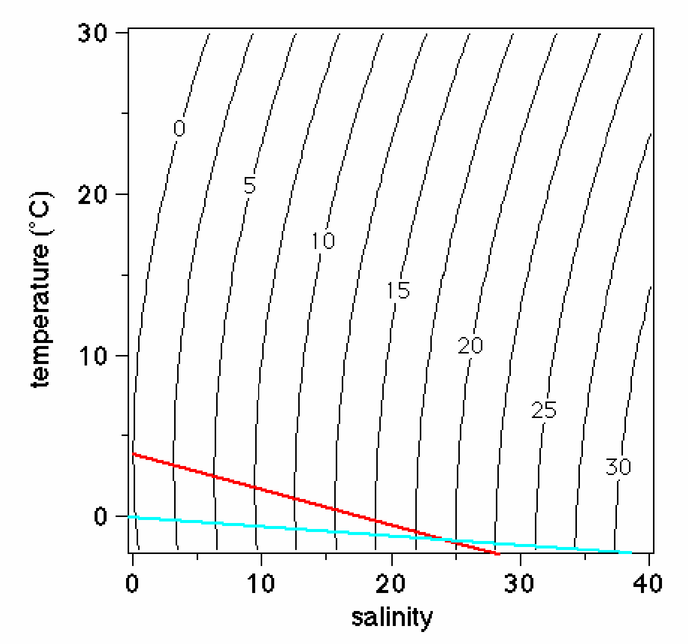


Fig. 3.4. Density σ_t as a function of temperature and salinity for the salinity range from freshwater to extreme oceanic salinities. The temperature of the density maximum is shown as a red line, the freezing point is shown as a light blue line.

A typical seawater density is thus $\sigma_t = 25$ (Figure 3.5).

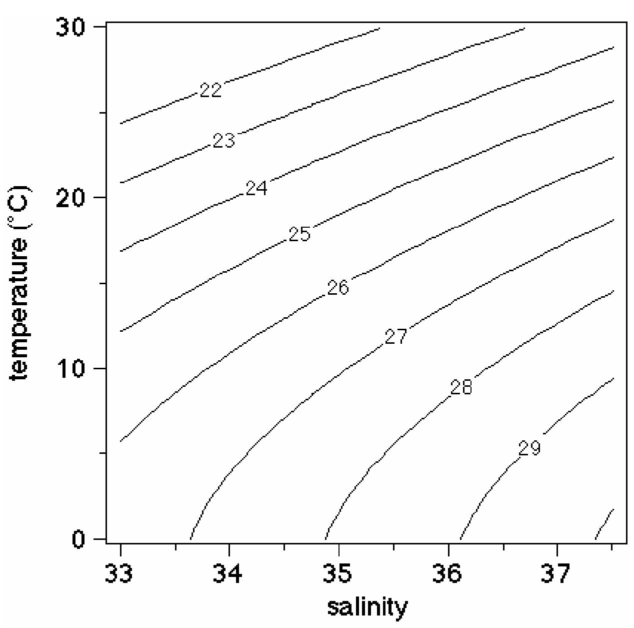


Fig. 3.5. A TS-diagram typical for oceanic applications. (Salinity is restricted to the range 33 to 37.5.)

A useful rule of thumb is that σ_t changes by the same amount if T changes by 1°C, S by 0.1, and p by the equivalent of a 50 m depth change.

Notice that the density maximum is above the freezing point for salinities below 24.7 but below the freezing point for salinities above 24.7. This affects the thermal convection:

- S < 24.7: The water cools until it reaches maximum density; then, when the surface water becomes lighter (ie after the density maximum has been passed) cooling is restricted to the wind-mixed layer, which eventually freezes over. The deep basins are filled with water of maximum density.
- S > 24.7: Convection always reaches the entire water body. Cooling is slowed down because a large amount of heat is stored in the water body. This is because the water reaches freezing point before the maximum density is attained.

Calculation based on Fofonoff, P. and R. C. Millard Jr (1983) Algorithms for computation of fundamental properties of seawater.Unesco Tech. Pap. in Mar. Sci. 44, 53 pp.

Lecture 4

The Global Oceanic Heat Budget

reviewed by: Alexandre Ganachaud

The oceanic heat budget consists of inputs and outputs. "Input" identifies a process through which the ocean gains heat, while "output" represents a heat loss to the ocean. A complete list of all inputs and outputs is as follows; + indicates input or heat gain, - signifies output or heat loss:
Primary **inputs** and **outputs**

- radiation from the sun (+)
- long-wave back radiation (-)
- direct heat transfer air/water (transfer of sensible heat) (-; + when from air to water)
- evaporative heat transfer (-; + when condensation; this situation occurs very rarely, mainly during sea fog conditions)
- advective heat transfer (currents, vertical convection, turbulence) (- or +); this effect cancels on the global scale or in closed basins

Secondary sources

- heat gain from chemical/biological processes (+)
- heat gain from the earth's interior and hydrothermal activity(+)
- heat gain from current friction (+)
- heat gain from radioactivity (+)

The contributions from secondary sources are negligible for most applications. The following discussion addresses only the primary inputs and outputs.

Heat Budget Inputs

Solar radiation

Energy from the sun at the outer limit of the atmosphere at normal incidence amounts to 2.00 cal cm^{-2} min^{-1} (the "solar constant"). Variations in the intensity of the incoming radiation are of regular and irregular character. Seasonally the solar radiation varies between zero and 1100 cal cm^{-2} day^{-1} at the poles and 800 - 900 cal cm^{-2} day^{-1} at the equator. Maximum interannual variations arise from the variation of the distance between the earth and the sun and amount to 3.34%; they can be predicted and explain the major climate changes over geological time.

In the modern literature the unit cal cm^{-2} day^{-1} (calories per square centimeter per day) has been replaced by the unit W m^{-2} (Watts per square meter). The translation between units is achieved by noting that 1 calorie (cal) = 4.184 Joules (J) and 1 Watt (W) = 1 Joule per second (J s^{-1}). This gives a conversion factor of 1 cal cm^{-2} day^{-1} = 0.484 W m^{-2}. In other words, a heat input of 1000 cal cm^{-2} day^{-1} converts roughly to 500 W m^{-2}.

Not all of the radiation received at the outer atmosphere is available to the ocean (Figure 4.1).

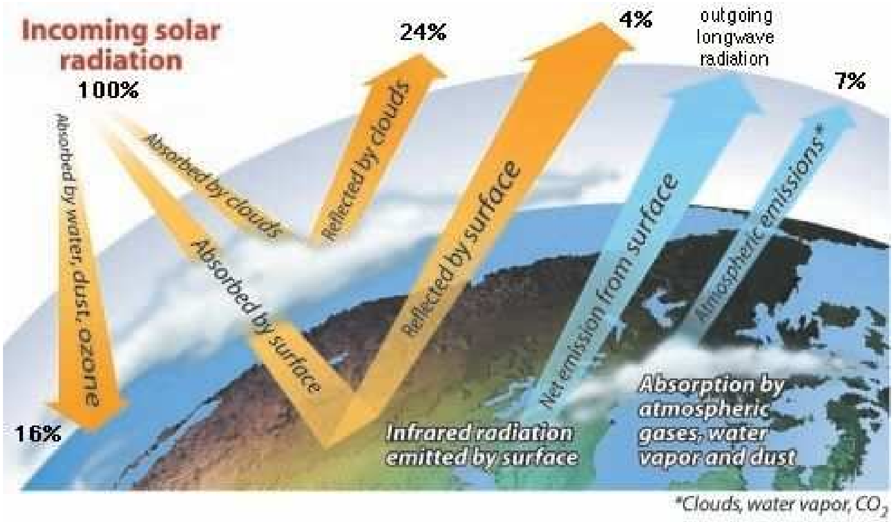


Figure 4.1. Absorption (16%), back radiation (7%) and reflection (24% from clouds, 4% from the earth's surface) of the incoming short-wave radiation in the atmosphere. Short-wave radiation reaches the ocean surface directly, as scattered clear sky radiation and as radiation scattered by clouds.
Adapted from Scripps Institution of Oceanography, *Explorations* vol. 10, no. 2, reproduced with permission.

If the incoming radiation is normalised to 100%, then

16% are absorbed in the atmosphere

24% are reflected by clouds

7% are radiated back to space from the atmosphere

4% are reflected from the earth's surface (mainly from the sea)

Thus, 35% returns into space, while 65% are available as energy. (The equivalent of 16% is stored in the atmosphere and therefore available eventually.)

The incoming radiation is emitted from the sun at ~6000 K (Kelvin, equivalent to the Celsius scale but with a shift in scale such that 0°C corresponds to 273 K). According to Wien's Law, maximum radiation occurs at a wavelength given by $\lambda = 2897\ T^{-1}$, where T is in degrees K and λ (lambda) the wavelength in micrometers. Maximum radiation from the sun therefore occurs in the wavelength range of visible light and peaks at 0.48 micrometers, which is in the blue range. It decays rapidly towards shorter wavelengths (in the ultraviolet or UV) and slowly towards longer wavelengths (in the infrared).

Solar energy received by the ocean varies irregularly with wavelength, as a result of absorbtion by water vapour and the various atmospheric gases, particularly oxygen and hydrocarbon. Absorption in the sea reduces the light level rapidly with depth (Figure 4.2).

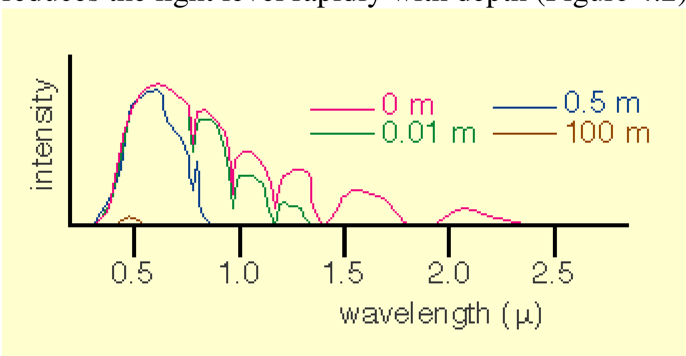


Figure 4.2. Spectral distribution of solar radiation received at different depths as a function of wavelength. The distinct minima in the incoming intensity at sea level (0 m) are caused by absorption from atmospheric gases (mainly water vapour, carbon dioxide and ozone).

At vertical light incidence (ie most favourable conditions),

73%	reaches	1 cm depth
44.5%	reaches	1 m depth
22.2%	reaches	10 m depth
0.53%	reaches	100 m depth
0.0062%	reaches	200 m depth

The minimum energy supply necessary to maintain photosynthesis is 0.003 cal cm^{-2} min^{-1}. Under optimum conditions (absolutely clear water) this amount is available at 220 m depth.

Heat Budget Outputs

Back Radiation

Some of the radiation received from the sun is radiated back from the ocean surface. The wavelength where most of this back-radiation occurs is again given by Wien's Law. As the temperature of the sea surface is much lower than that of the sun (~283 K), maximum back radiation is found at 10 micrometers, i.e. in the infrared or heat radiation.

According to the Stefan-Boltzman Law, the energy of the radiation is proportional to the fourth power of the absolute temperature (temperature expressed in K). Thus, daily or seasonal variations in the ocean's surface temperature have little effect on the back radiation energy, since these variations are small compared to the absolute temperature level.

Direct (Sensible) Heat Transfer Between Ocean and Atmosphere

On average, the ocean surface is about 0.8°C warmer than the air above it. Direct heat transfer (transfer of sensible heat) therefore occurs usually from water to air and constitutes a heat loss. Heat transfer in that direction is achieved much more easily than in the opposite direction for two reasons:

1. It takes much less energy to heat air than water. The energy needed to increase the temperature of a layer of water 1 cm thick by 1°C is sufficient to raise the temperature of a layer of air 31 m thick by the same amount.

2. Heat input into the atmosphere from below causes instability (through a reduction of density at the ground) which results in atmospheric convection and turbulent upward transport of heat. In contrast, heat input into the ocean from above increases stability (through a reduction of density at the surface) and prevents efficient heat penetration into the deep layers.

Evaporative Heat Transfer

51% of the heat input into the ocean is used for evaporation. In addition to the important contribution to the heat budget, evaporation - constituting a loss of water to the atmosphere - plays an important role in the mass budget, which will be discussed below.

Evaporation starts when the air is unsaturated with moisture. Warm air can retain much more moisture than cold air. As under normal conditions direct heat transfer is from the sea to the air (ie the air is normally heated from below), the normal situation is that the air is unsaturated with moisture and evaporation occurs. Condensation occurs where warm air is found over cold water. Such ocean areas are known and feared for the frequent occurrence of fog. Most of the energy released during condensation goes into the atmosphere, so the contribution of condensation to the oceanic heat budget is extremely small.

The heat budget is the balance between the terms discussed above. Normally, the first two terms are not considered separately; the difference **solar radiation minus oceanic back-radiation**, or net radiative heat gain, is used as the major input. The balance is then

net radiative heat gain - evaporative heat loss - direct heat loss = 0

This budget is closed if the world ocean is considered. If the budget is evaluated for limited ocean regions, the right-hand side is not usually zero but represents the heat transfer achieved by ocean currents. Figure 4.3 gives the heat budget of the North Atlantic Ocean as an example. On the global scale, Figure 4.4 shows the role of heat transport by currents in the heat budget in a zonal section from 60°N to 60°S.

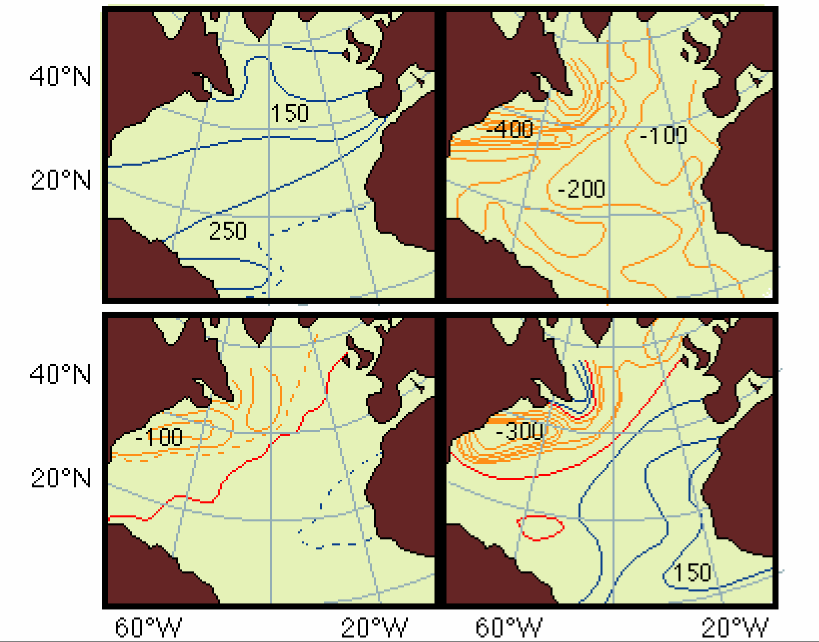


Figure 4.3. Heat budget of the North Atlantic Ocean. Contours are in cal cm^{-2}day^{-1}; contouring interval is 50 cal cm^{-2}day^{-1}, broken contours indicate intermediate 25 cal cm^{-2}day^{-1} contours. Oceanic heat gain is indicated by blue contours, oceanic heat loss by gold contours; the zero contour is red.
Top left: Net heat gain from solar radiation
Top right: Evaporative heat loss
Bottom left: Sensible heat loss
In the absence of currents and continents, the contours in these panels would be zonal (running east-west). Deviations are caused by heat transported in meridional (north-south running) ocean currents caused by land barriers. (This type of heat transfer is called advective heat transport.) Notice that over most of the ocean, evaporative heat loss nearly matches the net heat input. Sensible heat *loss* is large in the Gulf Stream region where warm water is moved from the tropics to temperate regions and is significantly warmer then the air; this also increases the evaporative heat loss. Sensible heat *gain* occurs on the east coast where cold water upwells to the surface, making the water at the surface colder than the air.
The lower right panel gives the sum of the other three panels. It shows large imbalances of the local heat budget, which have to be accounted for by transport of heat through ocean currents. This oceanic heat transport is therefore the inverse of the lower right panel. It is dominated by the transport of warm water in the Gulf Stream, i.e. large heat *gain* in the Gulf Stream region from water advected from the tropics.

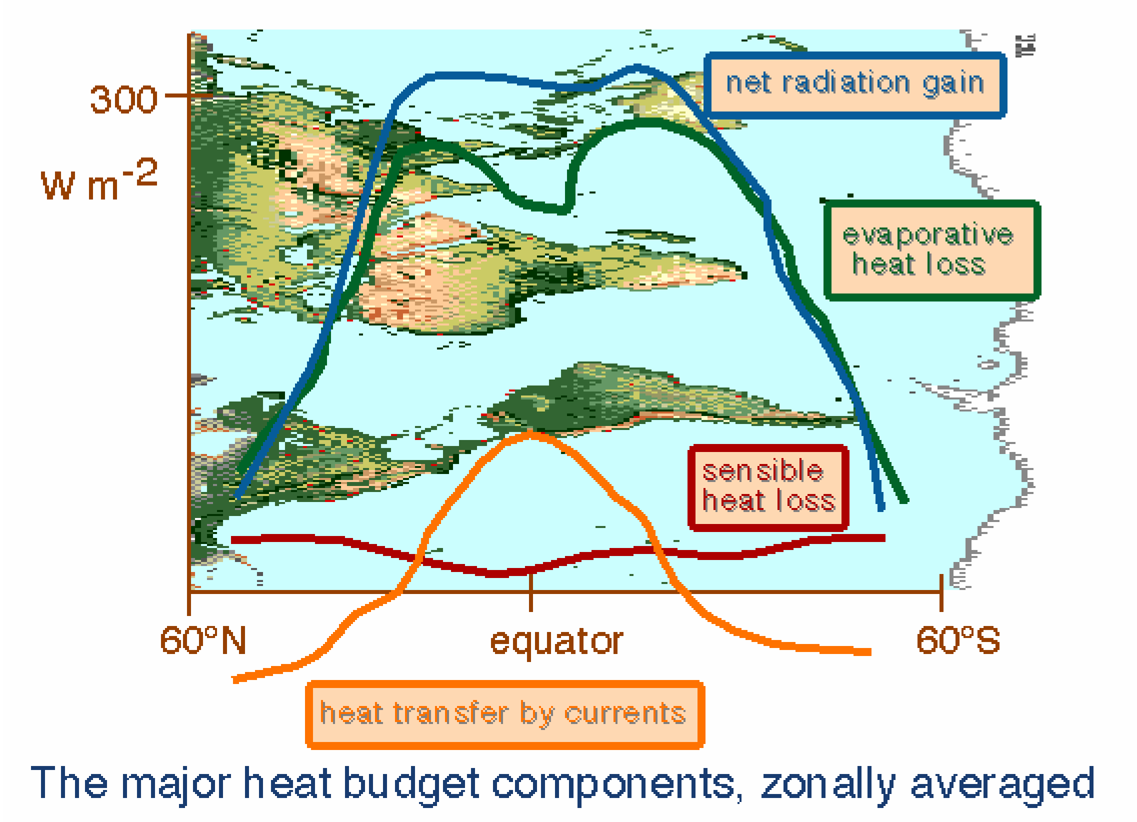


Figure 4.4. A qualitative sketch of the major heat budget components, zonally averaged over the oceans, as a function of latitude.

Net heat input decreases from the tropics towards the poles; it has a weak minimum near the equator because of heavy cloud cover in that region. The maximum of the evaporative heat loss in the subtropics is produced by atmospheric advection of dry air; the minimum in the tropics results from high moisture content of the tropical air. Sensible heat loss is small throughout. Currents remove heat from the tropics (this is a heat loss to the ocean - positive values) and deposit it in the subpolar regions (a heat gain - negative values).

The Oceanic Mass Budget

The mass budget involves the effects of evaporation and precipitation on the amount of water in the ocean. The effect on the total amount of water is significant only on a geological time scale. Its major importance for oceanographic applications lies in its influence on the salinity in the surface layer of the ocean.

The evaporation rate *E*, ie the loss of water due to evaporation over a given time span is proportional to the distribution of evaporative heat loss. The proportionality constant is known as the evaporation constant of water; it amounts to 585 cal g^{-1}.

Precipitation *P* has to be taken from observations. It is high in the Doldrums (just north of the equator) and at the Polar fronts (at about 50° latitude).

The distribution of sea surface salinity mirrors the distribution of *E-P* over large parts of the ocean (Figure 4.5).

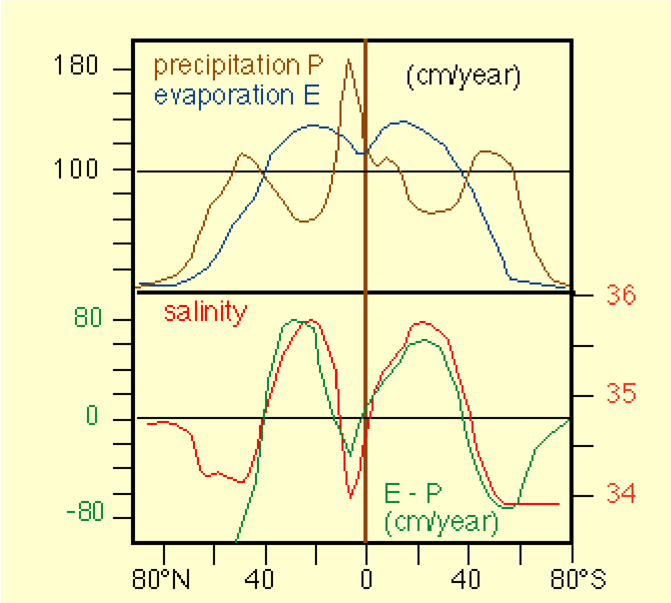


Figure 4.5. Annual mean distribution of evaporation E, precipitation P, the difference E - P (left scale), and sea surface salinity (right scale).

Deviations occur from river run-off. On a global scale, the balance is

Evaporation = $440 \cdot 10^3$ km^3 year^{-1}

Precipitation = $411 \cdot 10^3$ km^3 year^{-1}

River run-off = $29 \cdot 10^3$ km^3 year^{-1}

The melting and freezing of ice is balanced (except on the geological time scale). Most rivers are found in the northern hemisphere, so the proportionality between sea surface salinity and *E-P* is better over most of the southern hemisphere.

Note: Evaporation, precipitation and river run-off are expressed as volume per unit time. Modern oceanography uses more and more a unit called "sverdrup" (Sv), defined as 1 million cubic meters per second: 1 Sv = 10^6 m^3 s^{-1}. The conversion factor for km^3 year^{-1} to Sv is 1000 km^3 year^{-1} = 0.0317 Sv. This gives an evaporation of 14.0 Sv, a precipitation of 13.1 Sv and a total river run-off of 0.9 Sv.

Lecture 5

Distribution of temperature and salinity with depth; the density stratification

The discussion of the previous lectures concentrated on air/sea interaction processes and therefore addressed the distribution of temperature and salinity only in the surface layer, where regional and seasonal variations are large. However, most of the ocean is filled with water of relatively uniform temperature and salinity (Figure 5.1).

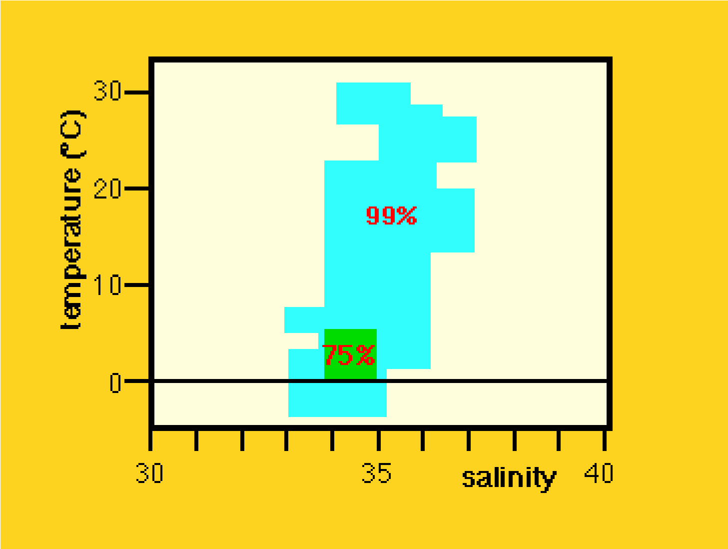


Figure 5.1. Volumetric temperature-salinity diagram of the world ocean. 75% of the ocean's water have a temperature and salinity within the green region, 99% have a temperature and salinity within the region coloured in cyan. The warm water outside the 75% region is confined to the upper 1000 m of the ocean.

If the surface temperature is very low, convection from cooling can reach deeper than the surface layer. This situation is encountered in the polar regions where cold water sinks to the bottom of the ocean. This process replenishes the deeper waters and is responsible for the currents below the upper kilometre of the ocean. Areas of deep winter convection are the Weddell Sea and the Ross Sea in the Southern Ocean and the Greenland Sea and the Labrador Sea in the Arctic region.

The average ocean temperature is 3.8°C; even at the equator the average temperature is as low as 4.9°C. The layer where the temperature changes rapidly with depth, which is found in the temperature range 8 - 15°C, is called the permanent thermocline. It is located at 150 - 400 m depth in the tropics and at 400 - 1000 m depth in the subtropics. Figure 5.2 shows the temperature and salinity distribution in a meridional section through the Pacific Ocean as an example.

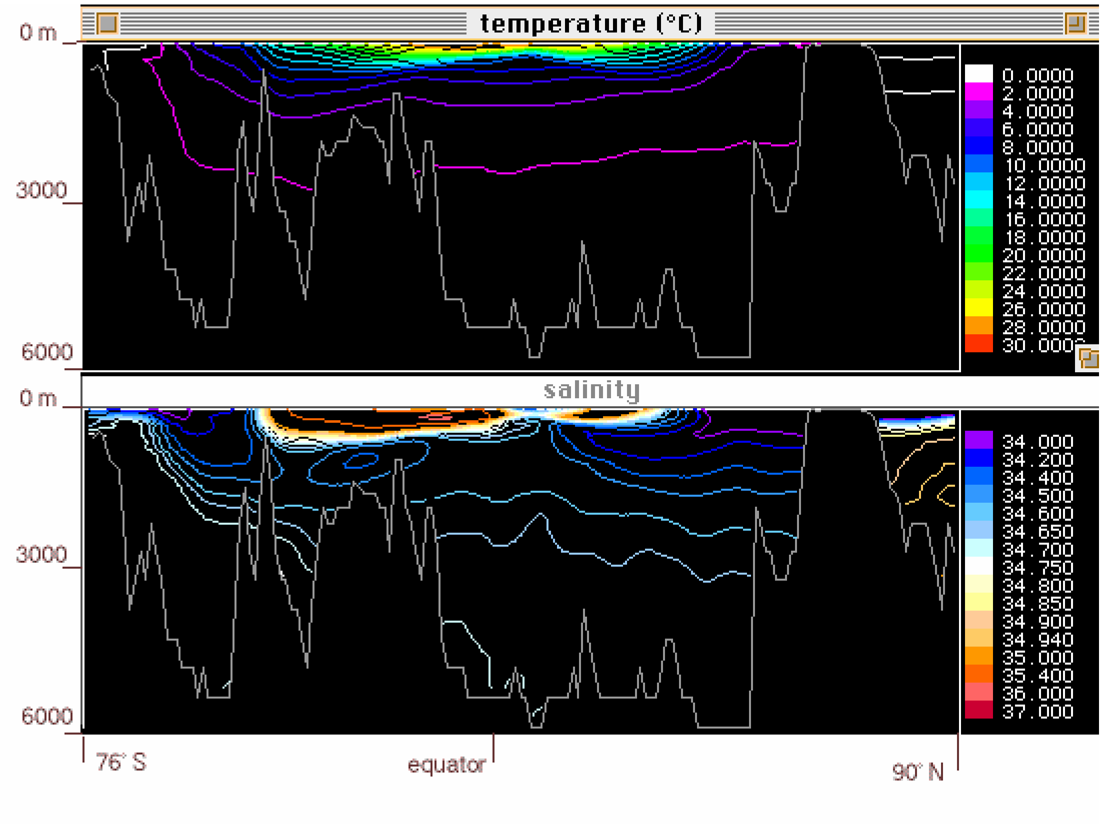


Figure 5.2. Temperature (top) and salinity (bottom) as functions of latitude and depth in the Pacific Ocean. (The image includes the Arctic Ocean on the extreme right.)
Note the uniformity of both properties below 1000 m depth; the temperature is in the range 0 - 4°C, the salinity is near 34.6 - 34.7.

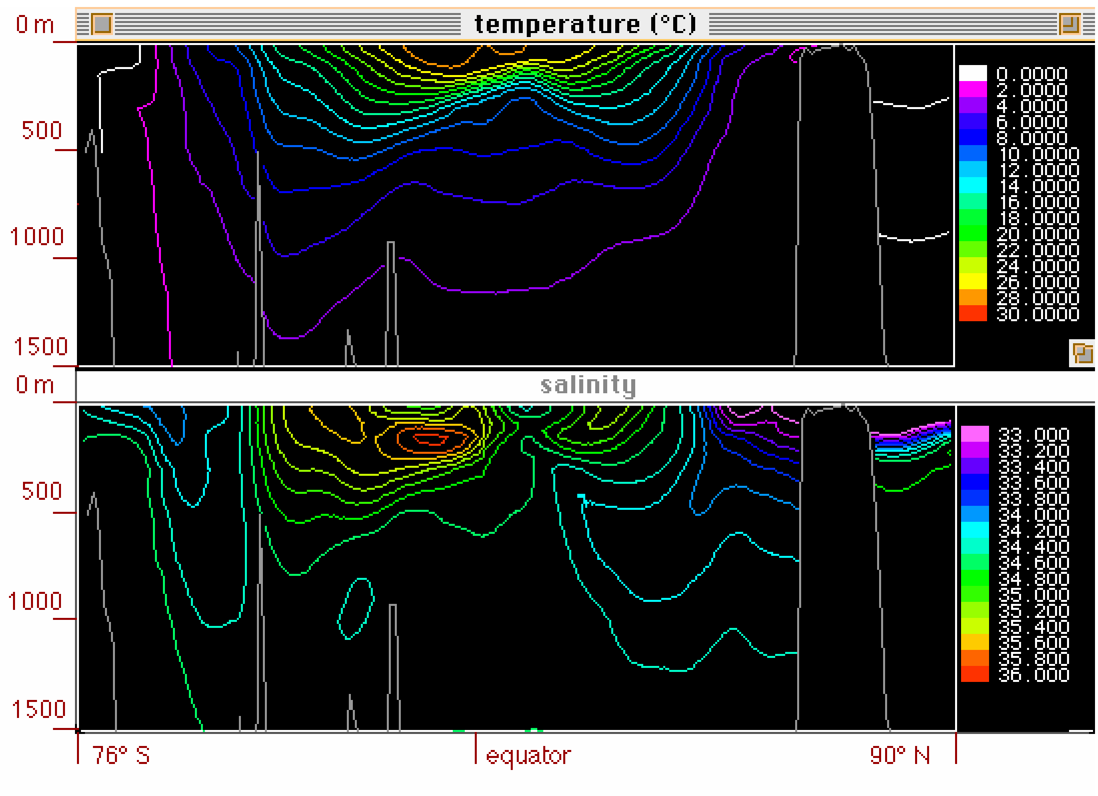


Figure 5.2a. Same as Figure 5.2, but upper 1500 m enlarged.

Notice the uniformity of both properties below 1000 m depth. Notice also that in many ocean regions, temperature and salinity both decrease with depth. A decrease in temperature results in an increase of density, so the temperature stratification produces a stable density stratification. A decrease in salinity, on the other hand, produces a density decrease. Taken on its own, the salinity stratification would therefore produce an unstable density stratification. In the ocean the effect of the temperature decrease is stronger than the effect of the salinity decrease, so the ocean is stably stratified.

In contrast to the subsurface temperature distribution, the subsurface salinity distribution shows intermediate minima. They are linked with water mass formation at the Polar Fronts where precipitation is high; details will be discussed later in the course. At very great depth, salinity increases again because the water near the ocean bottom originates from polar regions where it sinks during the winter; freezing during the process increases its salinity.

Acoustic Properties

Light and sound are two principal carriers of information used in human and animal communication. On land, sound is attenuated over much shorter distances than light, which is therefore the preferred choice for long-distance communication. The reverse situation is found in the ocean: While light does not penetrate very far in water, sound can travel over large distances and is therefore used for various purposes, such as depth sounding, communication, range finding and underwater measurement, by animals and humans alike. Detailed information on the speed of sound (ie the phase velocity of the sound waves) is essential for such applications.

The sound speed c is a function of temperature T,salinity S and pressure p and varies between 1400 m s^{-1} and 1600 m s^{-1}. In the open ocean it is influenced by the distribution of T and p but not much by S. It decreases with decreasing T, p and S. The combination of the variation of these three parameters with depth produces a vertical sound speed profile with a marked sound speed minimum at intermediate depth: Temperature decreases rapidly in the upper kilometre of the ocean and dominates the sound speed profile, i.e. c decreases with depth. In the deeper regions (below the top kilometre or so) the temperature change with depth is small and c is determined by the pressure increase with depth, ie c increases with depth. Vertical changes of salinity are too small to have an impact; but the average salinity determines whether c is low (if the average salinity is low) or high (if the average salinity is high) on average.

Figure 5.3 shows examples of sound speed profiles.

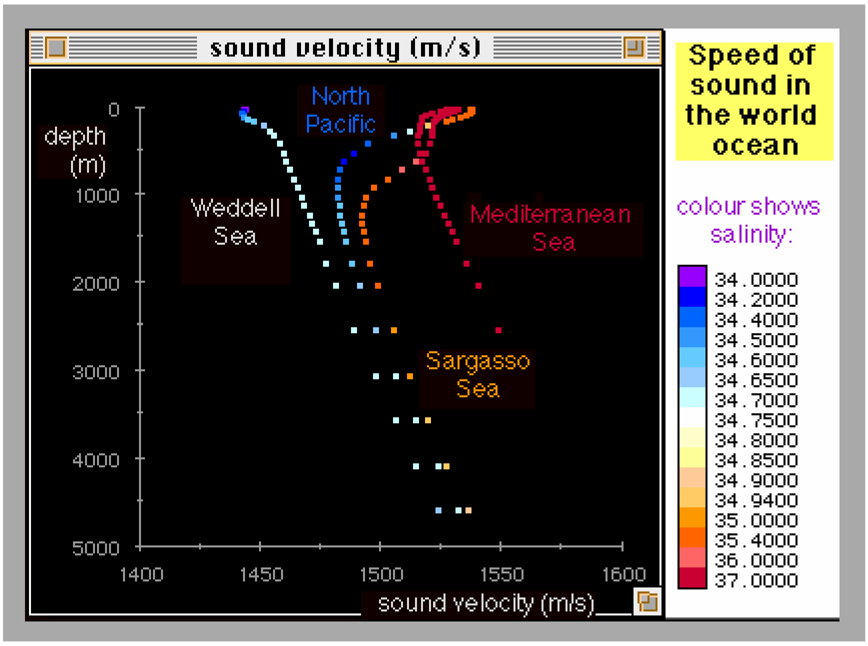


Figure 5.3. Sound speed *c* as a function of depth in various ocean regions.
Near the ocean surface, *c* decreases as a result of the temperature decrease with depth. At greater depth, temperature changes are small and *c* increases linearly with pressure (depth).
The effect of salinity on variations of *c* with depth are negligible, but salinity determines the overall magnitude of *c*: Salinity is low in the Pacific Ocean, higher in the Sargasso Sea (Atlantic Ocean) and highest in the Mediterranean Sea.
The Weddell Sea is an example of a polar region: Temperature is uniform over the entire depth range and extremely low, so *c* is also very low and shows only a vertical dependence on pressure (depth).

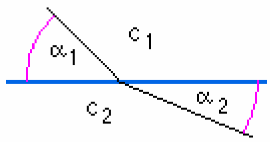
Note the curves for the Weddell Sea and the Mediterranean Sea: The Weddell Sea does not have a thermal stratification, hence no temperature effect on *c*. The Mediterranean Sea demonstrates the effect of salinity on *c*: The profile is similar to those of other tropical ocean regions, but the higher salinity of the Mediterranean Sea increases *c* at all levels.

Sound propagation

Sound propagates along rays (just as light does). Thus, the laws of geometrical optics are applicable to sound:

1. Sound travels along a straight path where the sound speed *c* is constant; it bends toward the region of lower *c* otherwise.
2. Different rays are independent of each other.
3. Sound paths are reversible.
4. The law of reflection (angle of incidence = angle of reflection) holds at the sea floor, surface, at objects and interfaces.
5. The law of refraction holds at interfaces:

$$\frac{\sin \alpha_1}{\sin \alpha_2} = \frac{c_1}{c_2}$$



As the stratification in the ocean is nearly horizontal, sound propagation in the vertical is practically along a straight path. This is the basis for echo sounding: The depth is known if the mean sound velocity is known. A first estimate is 1500 m s^{-1}; available tables list corrections for the various areas of the world ocean.

Figure 5.4 gives examples of horizontal sound paths.

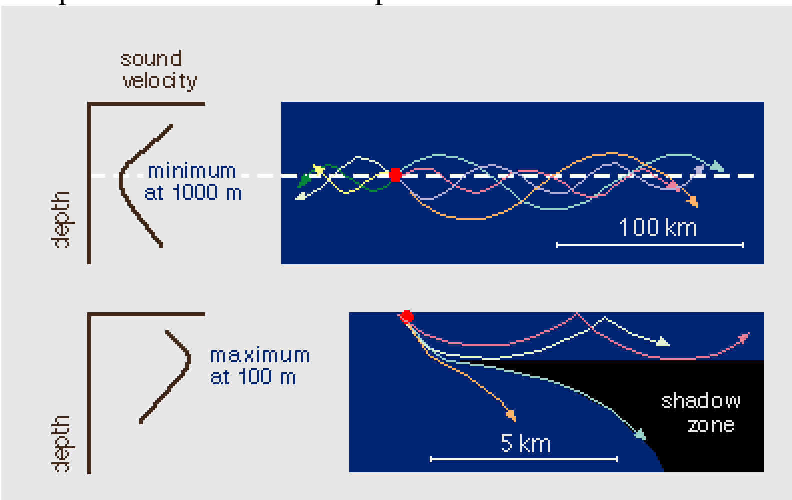


Figure 5.4. Horizontal sound paths in the SOFAR channel (top) and near the surface if the surface layer is well mixed (bottom). The diagram at the left gives the vertical sound velocity profile for each case; note the very different vertical scales. The red circle indicates the sound source.
The principle of sound propagation is that sound rays always bend towards the region of lower sound speed. This produces a sound channel near 1000 m depth (top) and a shadow zone below the surface mixed layer (bottom) to which sound from the source cannot penetrate. Since sound rays are reversible, this also means that sound produced in the shadow zone cannot be heard by a sound detector placed at the location of the red circle.

The first diagram shows sound propagation at the depth of the sound speed minimum (usually about 1000 m). Sound rays bend back towards the depth of minimum sound speed and travel at that depth over large distances (they can traverse entire oceans). This sound channel is known as the SOFAR (SOund Fixing And Ranging) channel. Before the introduction of the Global Positioning System (GPS) the SOFAR channel was used to locate ships and aircraft in distress, and for tracking floats (with two or more receivers) for the study of ocean currents. The second diagram shows a situation where a mixed layer of uniform temperature (typically about 100 m thick) is found on top of the normal temperature stratification. In this case sound speed increases below the surface due to the increase in pressure before the normal decrease due to temperature takes over. The resulting sound speed maximum at about 100 m depth creates a shadow zone, since all sound rays bend away from that depth.

Nutrients, oxygen and growth-limiting trace metals in the ocean

Justus von Liebig discovered what has become known as the "Minimum Law" of agriculture, that ecosystem productivity is limited by the nutrient which is exhausted first. On land the limiting element is either phosphorus, nitrogen or potassium (depending on soil type). In the ocean Liebig's Law indicates that the limiting elements should be

phosphorus	(as organic or anorganic phosphate)
nitrogen	(as nitrate, nitrite, and ammonia)

silicon (as silicate)

On land nutrients enter the soil by decomposition of dead organic matter. In the ocean nutrient uptake by plants (phytoplankton) occurs in the euphotic zone (the surface layer reached by sunlight) where photosynthesis takes place. Most nutrients are removed from the euphotic zone and transferred to the deeper ocean as dead organisms (detritus) sink to the ocean floor. In the deeper layers organic matter is remineralized, ie nutrients are brought back into solution. This process requires oxygen. Thus,

- the ocean cannot support highly productive ecosystems except where nutrients are returned to the euphotic zone from below (upwelling)
- nutrient concentrations usually increase with depth (Figure 5.5), while oxygen concentration (Figure 5.6) decreases. Departures from this rule are caused by advection of different water.

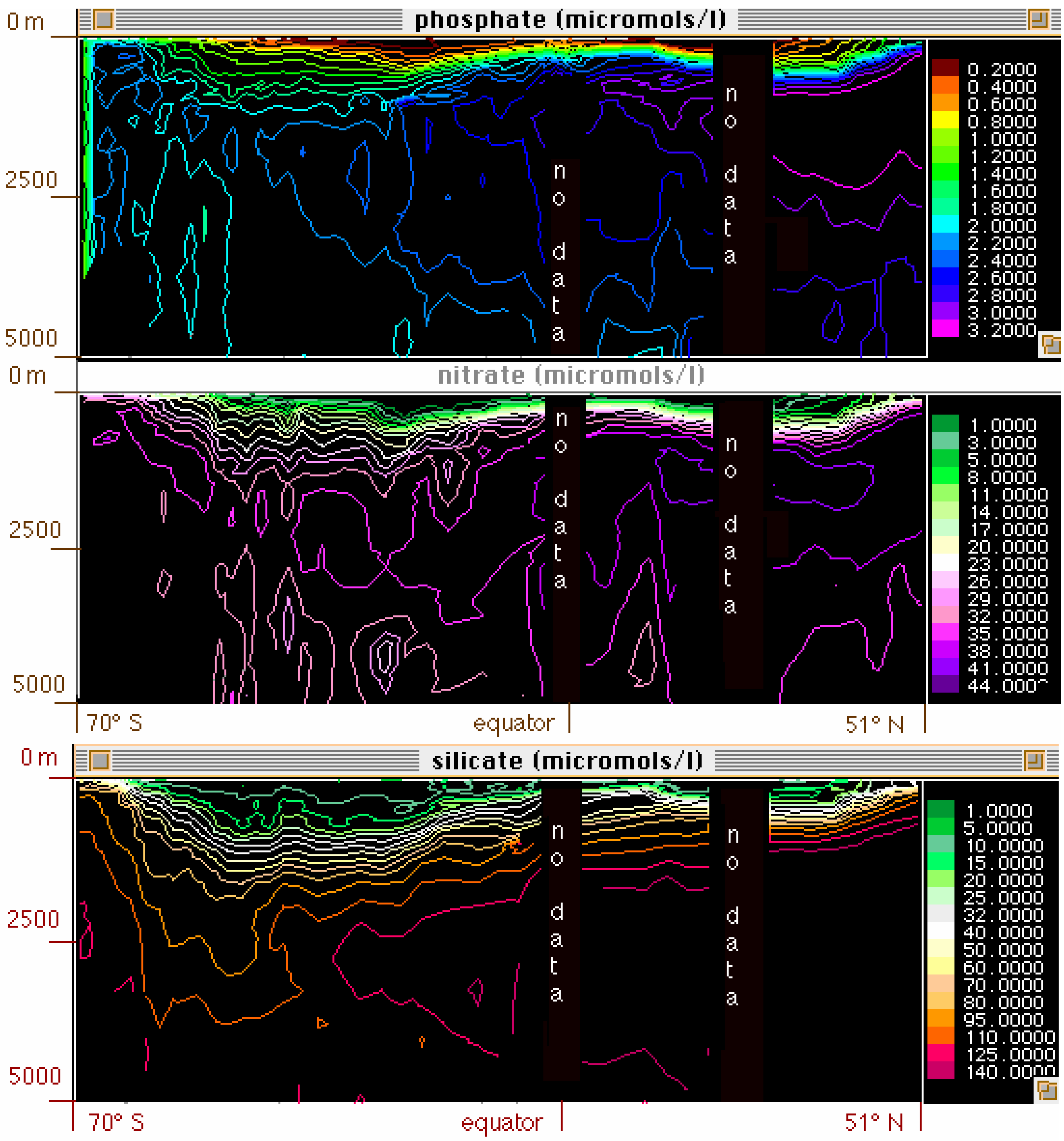


Figure 5.5. A meridional section of phosphate, nitrate and silicate concentration for the upper 5000 m of the Pacific Ocean along 170°W (about 100 km east of the date line). Note the large increase of nutrient concentration with depth; the deep ocean is a huge reservoir of stored nutrients.

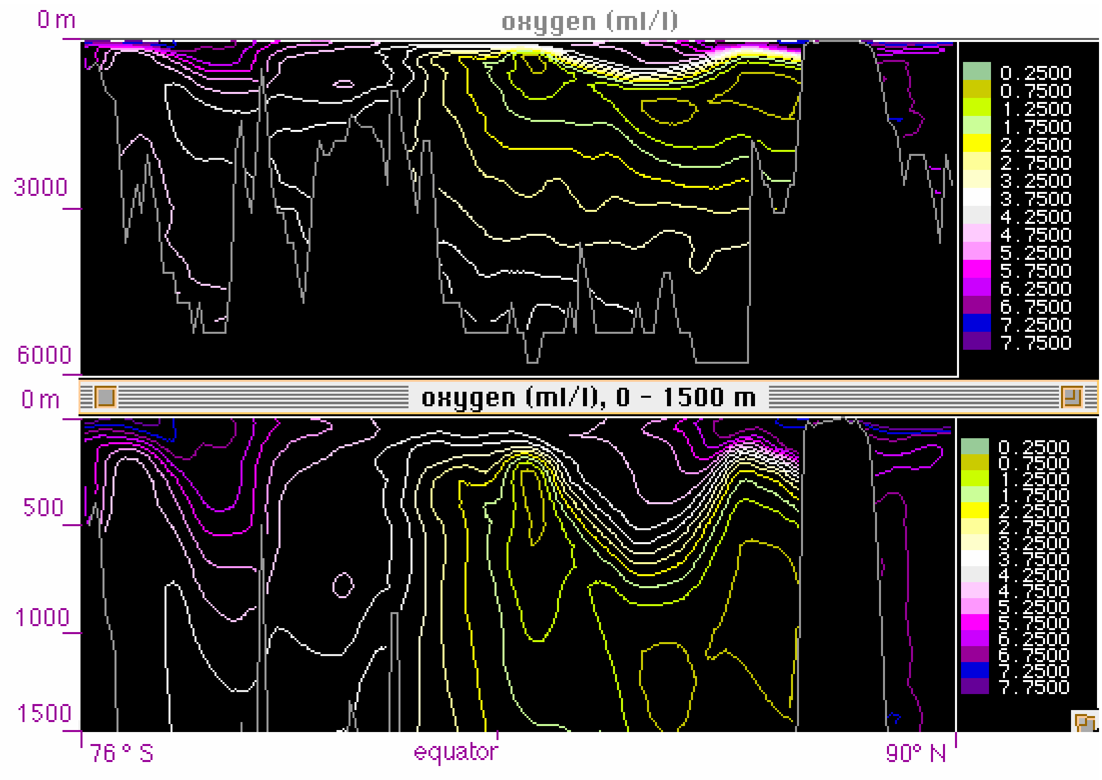


Figure 5.6. A meridional section of oxygen concentration for the Pacific Ocean along the date line. The lower figure is an enlargement of the upper 1500 m of the same section.
Note the decrease of oxygen concentration with depth. At the surface, oxygen concentration is above 7 ml/l in the polar regions and closer to 4.5 ml/l in the tropics. Below about 1000 m the concentration falls off to levels near 4 ml/l. The North Pacific Ocean is particularly low in oxygen at depth, with oxygen concentrations below 2 ml/l. Such extreme values are not typical for all deep ocean basins. The low oxygen concentratrations at depth are mostly due to oxygen uptake as nutrients are brought back into solution (remineralised); oxygen uptake by marine life is only of very minor importance.

Oxygen and nutrients are linked in a cycle of uptake and release, so a fixed ration of their concentrations is found in open ocean water:

AOU : C : N : P = 212 : 106 : 16 : 1 on atomic weight

= 109 : 41 : 7.2 : 1 in grams

AOU (apparent oxygen utilisation) = saturation concentration - observed concentration

C = carbon N = nitrogen P = phosphorus

The last three decades of the last century have seen great progress in the understanding of ocean chemistry, and it has now become clear that phosphate, nitrate and silicate are not the only growth-limiting nutrients in the ocean. In more than 40% of ocean regions biological growth is limited by the supply of iron (Fe). The reason for this difference between land based ecosystems and marine ecosystems is found in the early evolution of the earth.

As described in the Introduction lecture, the composition of the atmosphere is the result of the presence of life on Earth (compare the figure). The first life forms to develop (the prokaryotes, which are basically just molecules surrounded by a membrane and cell wall) found an atmosphere that consisted mainly of carbon dioxide (CO_2). They used the chemical elements available in the ocean for storage, transport and transfer of energy. Iron is one of the most abundant elements and became essential to many cellular functions.

The advent of photosynthesis in plants changed the relative distribution of C, O and Fe dramatically. As the oxygen level of the atmosphere increased, the oxygen was initially reduced by the available iron, creating vast deposits of iron oxide in the earth's crust. Eventually the supply of free iron was depleted, and the build-up of oxygen that allowed the evolution of higher life forms began. But primitive marine life still requires Fe for its cell functions, and this explains why in the ocean iron is an additional limiting element and in many situations the limiting factor. Field experiments have shown that oceanic productivity increases dramatically when iron is added to the euphotic zone.

Lecture 6

Aspects of Geophysical Fluid Dynamics

reviewed by: Ivan Lebedev, Yin Soong and Loren Lockwood

The equation which describes motion in the ocean is derived from Newton's Second Law, which expresses conservation of momentum in the form

force = mass times acceleration, or

F = *m* **a**

(Here, **bold** characters indicate vectors, characters in *italics* indicate scalars.)

In fluids this equation is expressed in terms of forces per unit mass **F' = F/** *m*, so that

F' = d**v** / dt,

where **v** = (u,v,w) is the velocity in component notation along the axes x,y,z with x pointing east, y north and z downward (**a** = d**v**/dt is the acceleration). If there is more than one force, Newton's Second Law applies to the sum of all forces involved. The law applies in an absolute co-ordinate system, ie a system which is either stationary or moving at constant speed. Co-ordinate systems in oceanography are usually defined with their origin somewhere on the earth's surface (for example at the north pole). They are therefore neither stationary nor moving at constant speed but rotate with the Earth. If Newton's Second Law is expressed in a rotating coordinate system, it must include an **apparent** or **virtual** force which takes care of the effect of rotation.

Classification of forces for oceanography

1. Forces generating currents

external forces: (exerted on the fluid boundaries)	(a) tangential stress (force exerted by wind)
	(b) effects of thermo-haline driving (surface cooling, evaporation etc)
	(Strictly speaking, surface cooling and evaporation are not forces as such but lead to changes of density which translate into changes of the pressure field.)
internal forces: (exerted on all water particles)	(c) Interior pressure field (pressure gradient)
	(d) Tidal forces

2. Forces retarding currents

(a)	Friction (diffusion of momentum)
(b)	Diffusion of density (not a force, but has the effect of changing the pressure gradient)

Forces 1a) and 1b) act only at the boundaries; mathematically, they determine the boundary conditions of the geophysical problem but do not enter the equation of motion. Force 2a) acts on all water particles and is therefore part of the equation of motion (it is one of the forces in Newton's sum of forces). Force 2b) is not a force acting on water motion directly but changes the

temperature and salinity fields and therefore the density; its effect is felt through the pressure field.

Newton's Second Law in oceanography ("Equation of Motion")

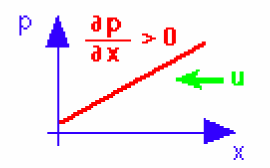
By taking the sum of all forces acting in the ocean Newton's Second Law takes the form

particle acceleration = - pressure gradient
+ Coriolis force
+ tidal force per unit mass
+ friction
+ gravity

The tidal force needs only be considered in tidal problems; it can be ignored in a discussion of the general oceanic circulation.

Gravity does not exert a horizontal force and thus cannot produce horizontal acceleration; it is important in movement which involves vertical motion (convection, waves).

Why the negative sign for the pressure gradient? Because the acceleration produced by a pressure gradient is directed opposite to the gradient, so the assocated water movement is "down the gradient":



pressure *p* increases with increasing distance *x* (to the right), the pressure gradient is positive, the acceleration is from high pressure to low pressure, the current *u* flows down the pressure gradient (to the left).

The Coriolis force (named after the French mathematician and engineer Gustave-Gaspard Coriolis, who described it in 1835) is an apparent force, that is, it is only apparent to an observer in a rotating frame of reference. To see this, consider a person standing on a merry-go-round, facing a ball thrown by a person from outside. To follow the ball the person would have to turn and therefore conclude that a force must be acting on the ball to deflect it from the shortest (straight) path. The person throwing the ball sees it follow a straight path and thus does not notice the force, and indeed the force does not exist for any person not on the merry-go-round. In oceanography currents are always expressed relative to the ocean floor - which rotates with the earth - and can therefore only be described correctly if the Coriolis force is taken into account in the balance of forces. The Coriolis force is proportional in magnitude to the flow speed and directed perpendicular to the direction of the flow. It acts to the left of the flow in the southern hemisphere and to the right in the northern hemisphere. A somewhat inaccurate but helpful way to see why the direction is different in the two hemispheres is related to the principle of conservation of angular momentum.

A water particle at rest at the equator carries angular momentum from the earth's rotation. When it is moved poleward it retains its angular momentum while its distance from the earth's axis is reduced. To conserve angular momentum it has to increase its rotation around the axis, just as ballet dancers increase their rate of rotation when pulling their arms towards their bodies (bringing them closer to their axis of rotation). The particle therefore starts spinning faster than the earth below it, ie it starts moving eastward. This results in a deflection from a straight path towards right in the northern hemisphere and towards left in the southern hemisphere. Likewise, a particle moving toward the equator from higher latitudes increases its distance from the axis of rotation and falls back in the rotation relative to the earth underneath; it starts moving westward, or again to the right in the northern and to the left in the southern hemisphere.

The effect of rotation on the apparent movement of objects can easily be demonstrated in laboratory experiments. You can see an example of such an experiment in the animation.

Inertial motion

If a water parcel is given some momentum (pushed) and then left alone, the only force acting on it is the Coriolis force. Newton's Second Law then states that the parcel will accelerate at a constant rate. Since the acceleration is produced by the Coriolis force it is directed perpendicular to the parcel's path; in other words, the acceleration takes the form of a constant change of direction. The result is that the parcel moves on a circle.

Inertial motion is very common in the ocean, where it is usually found superimposed on other movement (such as geostrophic or wind-driven flow discussed below). Figure 6.1 gives an example from the Baltic Sea.

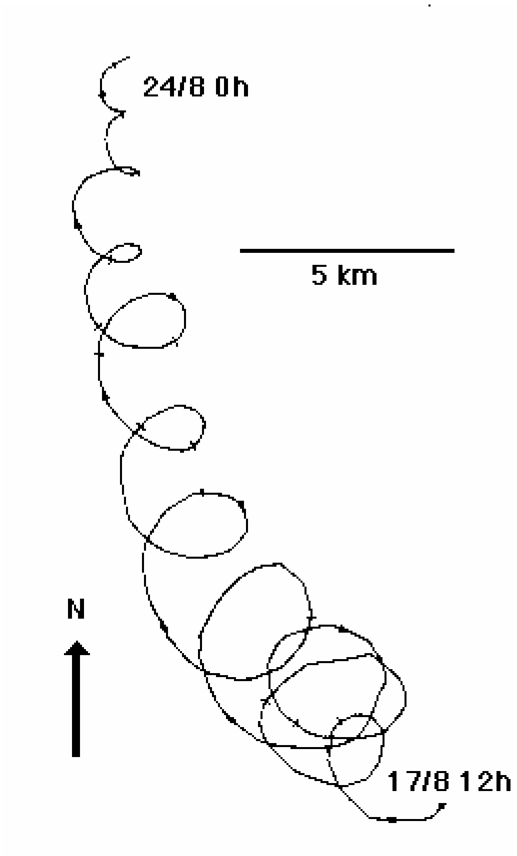


Figure 6.1. Inertial motion in the Baltic Sea, superimposed on a slow northwestward drift. Cross marks along the particle path indicate noon and midnight.
The water received a "push" (probably produced by a wind burst of several hours) some time after 17 August. It is then seen to move on circles. Friction reduces the inertial circles, until the inertial motion has nearly did away on 24 August.

after T. Gustafson and B. Kullenberg: "Untersuchungen von Trägheitsströmungen in der Ostsee." Svens. Hydrogr. Biol. Skr. Ny. Ser. Hydrogr. 13 (1936), 1 - 28.

Geostrophic flow

In the ocean interior, ie below about 100 m depth and about 100 km away from coastlines, frictional forces can be neglected. The steady state circulation is then determined by the balance between the pressure gradient force and the Coriolis force. This balance is known as **geostrophic flow**. In geostrophic flow particles move along isobars (contours of constant pressure), with high pressure on their **left** in the **southern** hemisphere, on their **right** in the **northern** hemisphere.

Since the pressure at any depth is determined by the weight of the water above, high and low pressure are equivalent to high sea level and low sea level. Geostrophic flow is therefore related to the shape of the sea surface.

The Coriolis force and the pressure gradient force act on all water particles. Geostrophic flow is therefore part of the oceanic current field at all depths and locations. Below about 100 m depth and about 100 km away from coastlines all currents are geostrophic; closer to the surface and to boundaries they are modified by additional forces.

Figure 6.2 shows an example of geostrophic flow in the equatorial current system. Note that the variations in sea level are only of the order 0.2 - 0.4 m. These small variations are impossible to verify in the open ocean. They have, however, been varified in narrow straits, where a reversal of the current flowing through the strait produces a reversal of the sea surface tilt across the strait. The tilt can be measured with sea level gauges on both sides.

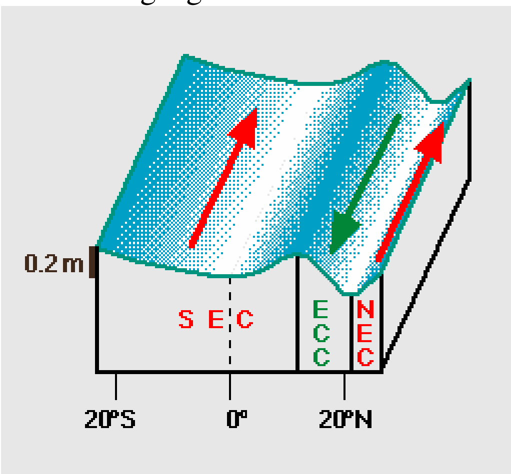


Figure 6.2. A sketch of the surface topography across the equatorial current system. NEC: North Equatorial Current, ECC: Equatorial Countercurrent, SEC: South Equatorial Current. The sea surface slopes up towards the north across the NEC, producing high pressure on the right of the current (which is in the northern hemisphere); it slopes down towards the north across the ECC, again producing high pressure on the right of the current (looking downstream). The SEC flows in the same direction as the NEC but is found on both sides of the equator, so the sea surface slopes up to the north, in the same way as it does across the NEC, where the SEC is flowing in the northern hemisphere; it slopes up to the south across the SEC in the southern hemisphere, producing high pressure on the left of the current. The variation of sea level necessary to maintain the pressure distribution is of the order of 0.2 - 0.4 m and cannot be seen when looking at the ocean. It has been verified through measurements of the shape of the ocean surface from satellites.

Another important aspect of geostrophic flow relates to the circulation around eddies. Figure 6.3 shows the principle. It applies equally to oceanic and atmospheric eddies and explains why a high pressure system to the east of Adelaide in South Australia brings northerly winds.

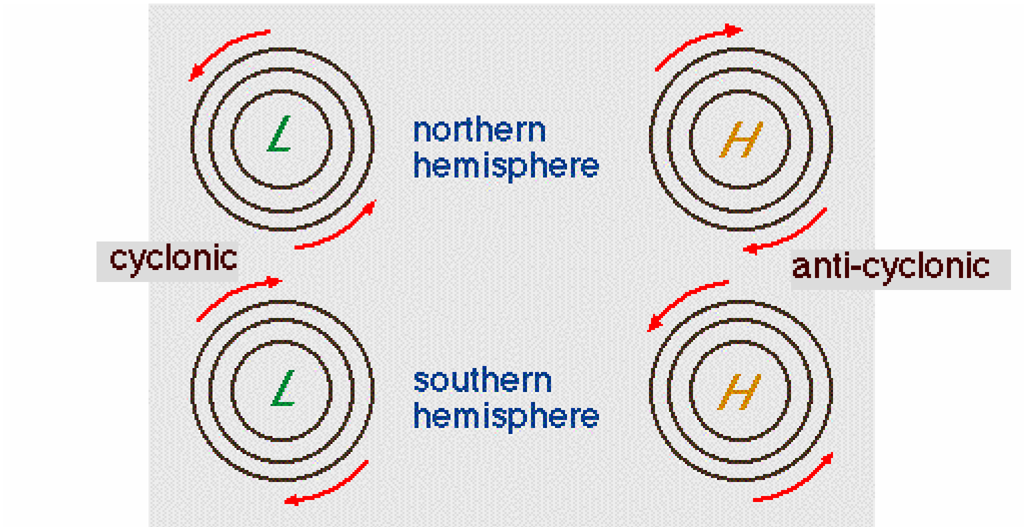


Figure 6.3. Current direction around high and low pressure centres in the northern and southern hemisphere. Circulation around centres of **low** pressure is called **cyclonic**, circulation around centres of **high** pressure is called **anti-cyclonic** (irrespective of hemisphere).
An alternative pair of terms not used very frequently refers to movement of the sun:

- contra-solem (literally "against the sun") for cyclonic,
- cum-sole (literally "with the sun") for anti-cyclonic.

Again, these terms are independent of hemisphere.
The terms "clockwise" and "anti-clockwise" are often used but are not independent of hemisphere:

- cyclonic motion is anti-clockwise in the northern hemisphere, clockwise in the southern hemisphere,
- anti-cyclonic motion is clockwise in the northern hemisphere, anti-clockwise in the southern hemisphere.

The Ekman Layer

Currents in the upper 150 m or so of the ocean are directly affected by the wind, ie transfer of momentum from the atmosphere to the ocean. The balance of forces therefore involves frictional forces, which cause a departure from geostrophic flow; water moves across isobars from areas of high pressure to areas of low pressure. The layer in which the flow is non-geostrophic is known as the Ekman layer.

The direction of water movement in the Ekman layer varies with depth. The details are complicated (Figure 6.4); but when only the steady state is considered, an important result is that the net (ie vertically averaged) transport in the Ekman layer is directed perpendicular to the wind direction, to the **left** in the **southern** hemisphere and to the **right** in the **northern** hemisphere.

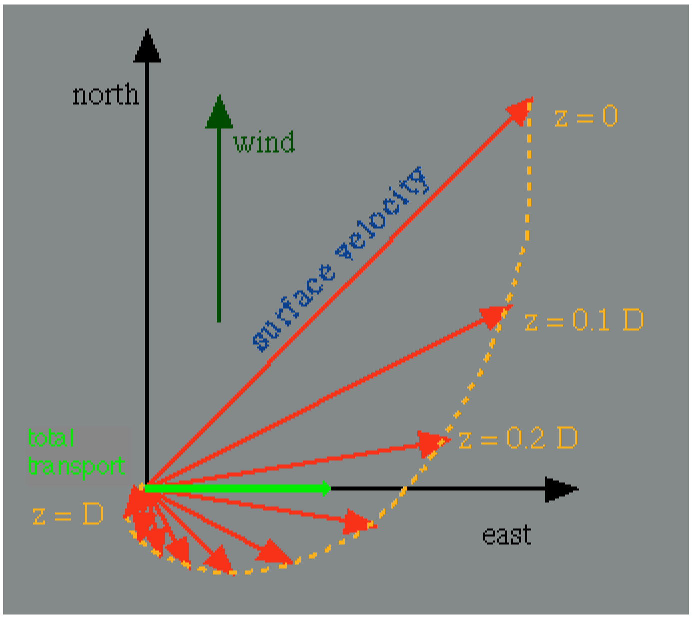


Figure 6.4. Variation of current in the Ekman layer with depth. D is the thickness of the Ekman layer. The current is shown from the surface (z=0) to the bottom of the Ekman layer (z=D) in depth increments of 1/10 of the Ekman layer thickness.
The current is 45° to the right of the wind at the surface and rotates clockwise as it decreases with depth. The net effect of the current distribution is water transport at right angle to the wind (ie to the east in this example.
The example assumes a southerly wind in the northern hemisphere. In the southern hemisphere the graph is the mirror image with respect to the vertical axis.
If the wind blows from a different direction the structure aligns with the new wind direction, such that the surface current is always at 45° from the wind direction.

Upwelling

Upwelling is the process of vertical water movement to the surface of the ocean. Coastal and equatorial upwelling are responses to prevailing winds and provide direct evidence of Ekman transport dynamics. A third type of upwelling is not directly wind-related; it occurs in the Southern Ocean and is an element of the global conveyor belt.

1. **Coastal upwelling**: Along the eastern coasts of the Pacific and Atlantic Oceans the Trade Winds blow nearly parallel to the coast towards the Doldrums. The Ekman transport is therefore directed offshore, forcing water up from below (usually from 200 - 400 m depth; Figure 6.5).
2. **Equatorial Upwelling**: In the Pacific and Atlantic Oceans the Doldrums are located at 5°N, so the southern hemisphere Trade Winds are present on either side of the equator. The Ekman layer transport is directed to the south in the southern hemisphere, to the north in the northern hemisphere. This causes a surface divergence at the equator and forces water to upwell (from about 150 - 200 m).
3. **Upwelling in the Southern Ocean**: North Atlantic Deep Water reaches the Southern Ocean in a broad flow in the depth range 1000 - 4000 m. It rises to within 200 m of the surface, to enter the circulation of the upper layers. This rise over more than 2000 m is the deepest upwelling process in the World Ocean.

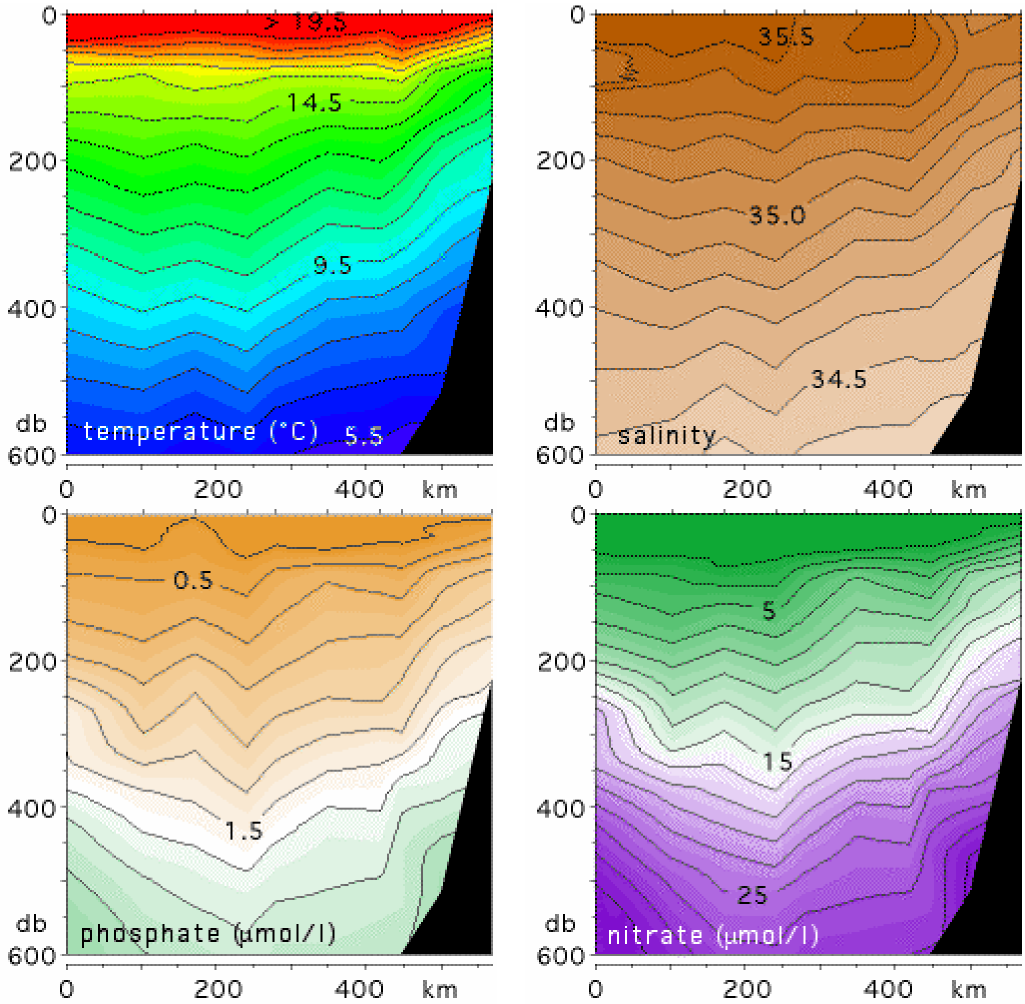


Figure 6.5. Water property sections in a coastal upwelling region, indicating upward water movement within about 200 km from the coast. (This particular example comes from the Benguela Current upwelling region, off the coast of Namibia.) The coast is on the right, outside the graphs; the edge of the shelf can just be seen rising to about 200 m depth at the right of each graph.

Note how all contours rise towards the surface as the coast is approached; they rise steeply in the last 200 km. On the shelf the water is colder, less saline and richer in nutrients as a result of upwelling.

Lecture 7

Thermohaline processes; water mass formation; the seasonal thermocline

reviewed by: Alexandre Ganachaud

In most ocean regions the wind-driven circulation, which was the focus of discussion so far, does not reach below the upper kilometre of the ocean. Water renewal below that depth is achieved by currents which are driven by density differences produced by temperature (thermal) and salinity (haline) effects. The associated circulation is therefore referred to as **the thermohaline circulation.** Since these movements are mostly very sluggish, it is often impractible to use current meters to measure them directly; they are usually deduced from the distribution of water properties and the application of geostrophy.

The driving force for the thermohaline circulation is water mass formation. Water masses with well-defined temperature and salinity characteristics are created by surface processes in specific locations; they then sink and mix slowly with other water masses as they move along. The two main processes of water mass formation are deep convection and subduction. Both are linked to the dynamics of the mixed layer at the surface of the ocean; so it is necessary to discuss thermohaline aspects of the upper ocean first.

Oceanographers refer to the surface layer with uniform hydrographic properties as the **surface mixed layer.** This layer is an essential element of heat and freshwater transfer between the atmosphere and the ocean. It usually occupies the uppermost 50 - 150 m or so but can reach much deeper in winter when cooling at the sea surface produces convective overturning of water, releasing heat stored in the ocean to the atmosphere. During spring and summer the mixed layer absorbs heat, moderating the earth's seasonal temperature extremes by storing heat until the following autumn and winter, and the deep mixed layer from the previous winter is covered by a shallow layer of warm, light water. During this time mixing is achieved by the action of wind waves, which cannot reach much deeper than a few tens of meters. Below the layer of active mixing is a zone of rapid transition, where (in most situations) temperature decreases rapidly with depth. This transition layer is called the **seasonal thermocline**. Being the bottom of the surface mixed layer, it is shallow in spring and summer, deep in autumn, and disappears in winter, when heat loss at the surface produces instability and the resulting convection mixes the water column to greater depth (Figure 7.1). In the tropics winter cooling is not strong enough to destroy the seasonal thermocline, and a shallow feature sometimes called the tropical thermocline is maintained throughout the year.

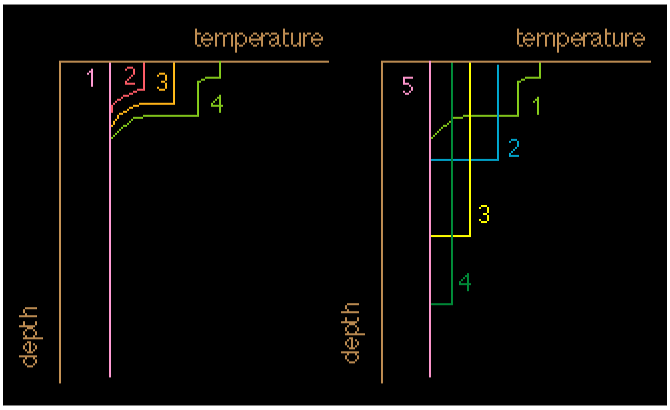


Figure 7.1. Development of the seasonal thermocline during the year.
Left: warming cycle. It starts with vertically homogeneous conditions (1). Heating at the surface warms the water; the heat is stirred into a mixed layer by wind mixing (2). This continues for several months (3). In mid-summer winds are often weaker than during spring, wind mixing does not reach quite so deep, and the mixed layer may consist of two or more homothermal layers (4).
Right: the cooling cycle. It starts at the end of summer (1, which is identical to 4 of the warming cycle). Cooling at the surface leads to instability and vertical overturn. This progressively deepens the mixed layer (2-4), until it disappears in winter (5).
Numbers can be approximately taken as successive months, with the following association:

	southern hemisphere		northern hemisphere	
number in graph	warming cycle	cooling cycle	warming cycle	cooling cycle
1	August	December	February	June
2	October	February	April	August
3	November	March	May	September
4	December	April	June	October
5		July		January

The depth range from below the seasonal thermocline to about 1000 m is known as the **permanent** or **oceanic thermocline**. It is the transition zone from the warm waters of the surface layer to the cold waters of great oceanic depth The temperature at the upper limit of the permanent thermocline depends on latitude, reaching from well above 20°C in the tropics to just above 15°C in temperate regions; at the lower limit temperatures are rather uniform around 4 - 6°C depending on the particular ocean.

Below the surface layer which is in permanent contact with the atmosphere, temperature and salinity are conservative properties, ie they can only be changed by mixing and advection. All other properties of sea water such as oxygen, nutrients etc. are affected by biological and chemical processes and therefore non- conservative. Water masses can therefore be identified by their temperature-salinity (T-S) combinations (Figure 7.2).

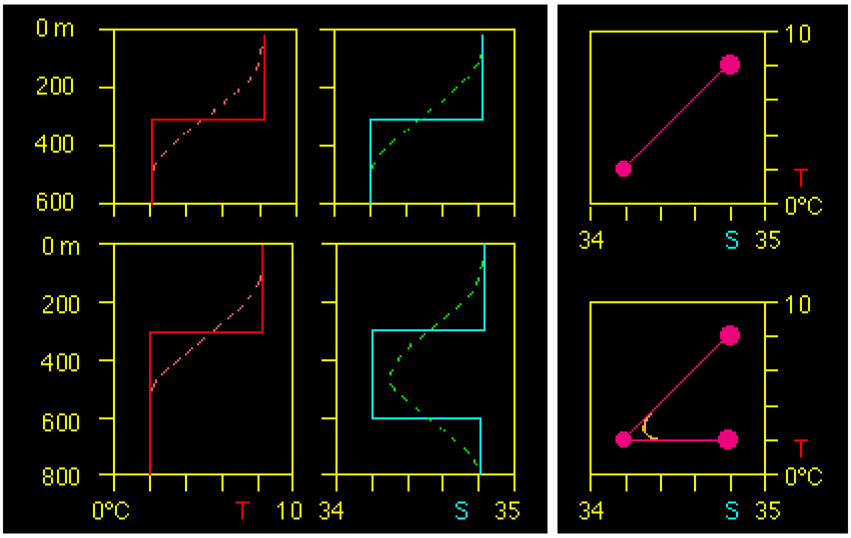


Figure 7.2. Examples of temperature-salinity (TS)-diagrams. The diagrams on the left show the distribution of temperature (red) and salinity (cyan) with depth; the diagrams on the right show the corresponding TS-diagrams. Top: layering of a warm and saline water mass found at 0 - 300 m depth above a cold and fresh water mass found at 300 - 600 m. The full lines show the situation before mixing, the broken lines after mixing. The TS-diagram shows the two water masses as TS points. Before mixing only the two points are seen in the TS-diagram. Mixing connects the two TS-points by a straight line.
Bottom: layering of three water masses (intrusion of a low salinity water mass at 300 - 600 m depth). Again, the full lines show the situation before mixing, the broken lines after mixing. The TS- diagram shows two mixing lines; the erosion of the intermediate salinity minimum by mixing is seen by the departure of the broken curve from the original water mass point.

Water mass formation by deep convection occurs in regions with little density stratification (ie mostly in polar and subpolar regions). When the water in the mixed layer gets denser than the water below, it sinks to great depth, in some regions to the ocean floor. The density increase can be achieved by cooling or an increase in salinity (either through evaporation or through brine concentration during freezing) or both.

Water mass formation by subduction occurs mainly in the subtropics. Water from the bottom of the mixed layer is pumped downward through a convergence in the Ekman transport and sinks slowly along surfaces of constant density (Figure 7.3).

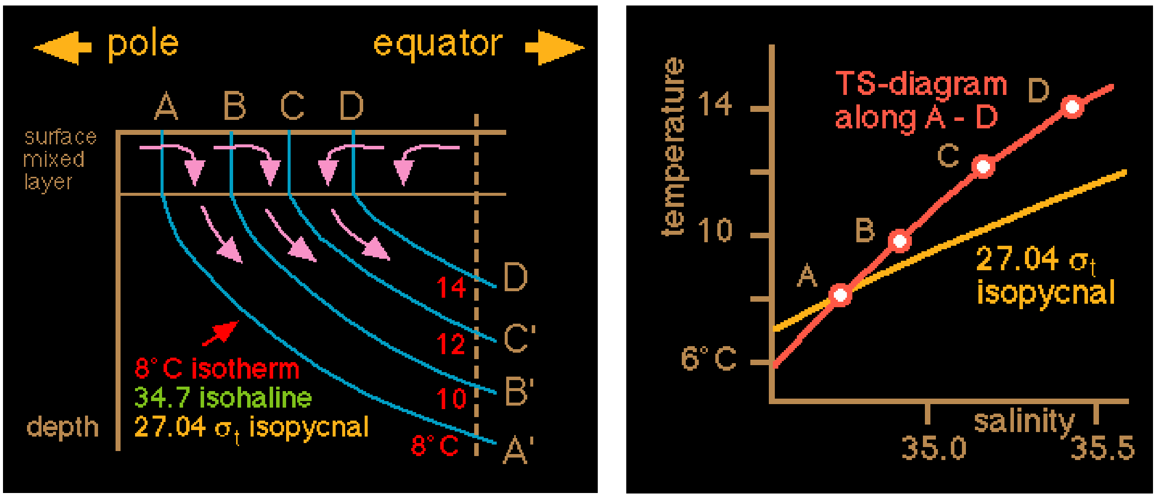


Figure 7.3. Sketch of water mass formation by subduction. First diagram: Convergence in the Ekman layer (surface mixed layer) forces water downward, where it moves along surfaces of constant density. The 27.04 σ_t surface, given by the TS-combination 8°C and 34.7 salinity, is identified.
Second diagram: A TS- diagram along the surface through stations A -> D is identical to a TS-diagram taken vertically along depths A´ - D´.

Figure 7.4 gives a summary of water masses in the world ocean.

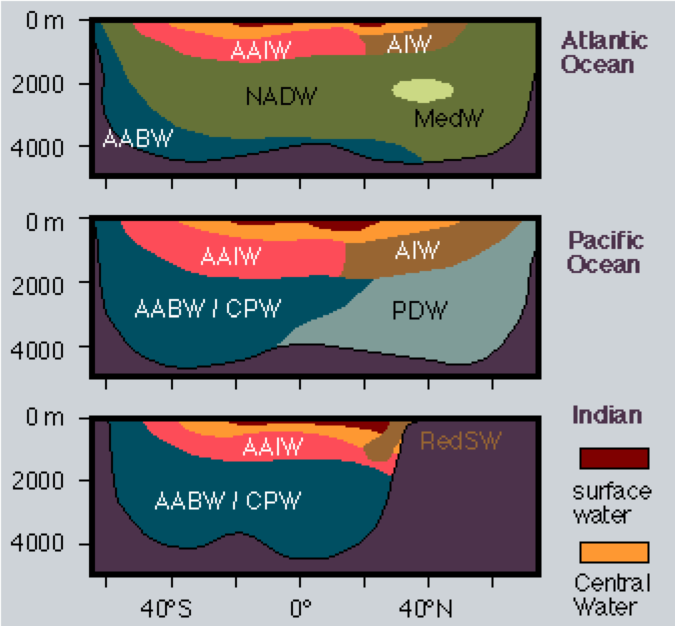


Figure 7.4. Sketch of the water mass distribution in the world ocean. AABW: Antarctic Bottom Water, CPW: Circumpolar Water, NADW: North Atlantic Deep Water, PDW: Pacific Deep Water, AAIW: Antarctic Intermediate Water, AIW: Arctic Intermediate Water, MedW: Mediterranean Water, RedSW: Red Sea Water, gold: Central Water, brown: surface water.

Antarctic Bottom Water is formed mainly in the Weddell and Ross Seas by deep convection and fills all ocean basins below 4000 m depth; in the Pacific and Indian Oceans it is mixed with North Atlantic Deep Water, the mixture being known as Circumpolar Water. **North Atlantic Deep Water** is the product of a process that involves deep convection in the Arctic Ocean, the Greenland Sea and the Labrador Sea. Most **Antarctic Intermediate Water** is formed by deep convection east of southern Chile and west of southern Argentina and spreads into all oceans with the Circumpolar Current. Intermediate Water in the northern hemipshere may be formed by convection or subduction. **Central Water**, the water of the permanent thermocline, is formed by subduction in the subtropics. Mediterranean and Red Sea waters are intrusions of high temperature, high salinity waters from two mediterranean seas (see the discussion of mediterranean seas below).

It is worth stressing that the picture developed here is of a very schematic nature. The real ocean is a fluid in turbulent motion full of eddies, fronts and other instabilities. It should also be kept in mind that significant zonal (east-west directed) movement occurs in every ocean basin and that the schematic distribution shown in Figure 7.4 cannot depict the variations that occur from the eastern to the western coasts. However, as a summary of the principal features of the water masses in the world ocean it is correct and adequate.

A summary of the TS characteristics of all water masses is given in Figure 7.5.

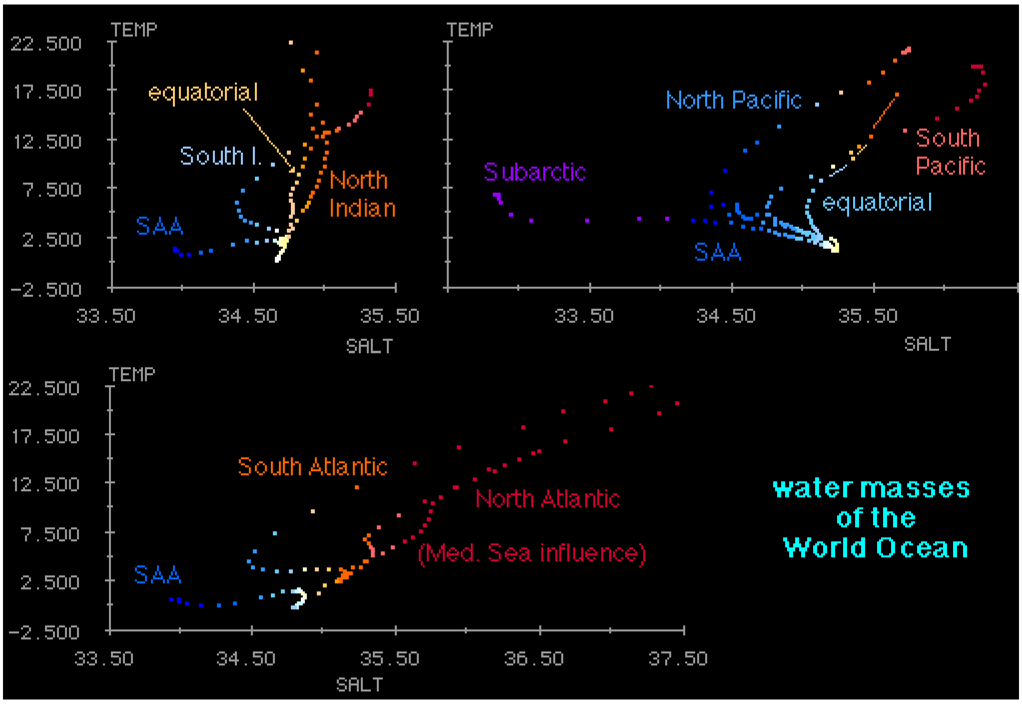


Figure 7.5. TS-diagrams of the Indian, Pacific and Atlantic Oceans showing the TS properties of the major water masses. Colour indicates salinity; the information is the same as given on the horizontal axis but helps to compare the three oceans.
All TS-diagrams "fan out" from a point below 0°C and a salinity of about 35.6 (white). This is Antarctic Bottom Water, found in all oceans. At higher temperatures the shape of the TS-diagrams (ie the TS-properties of the ocean) depend on the ocean region. In the Subantarctic zone (SAA) (south of 45°S) temperatures remain low and salinities decrease. The same is seen in the Subartic zone (north of 45°N). The influence of high salinity inflow from the Mediterranean Sea is seen by higher salinities near 8 - 10°C in the North Atlantic Ocean. The influence of high salinity outflow from the Red Sea raises the salinities in the North Indian Ocean.

Antarctic Bottom Water is represented by a single TS point (in the white region of the salinity scale). Antarctic Intermediate Water also has its own TS point but is usually only seen as a salinity minimum in the TS-curve; the minimum is slowly eroded by mixing as the Intermediate Water progresses northward (Figure 7.6). The Central Waters are represented by TS curves rather than points (compare Figure 7.3).

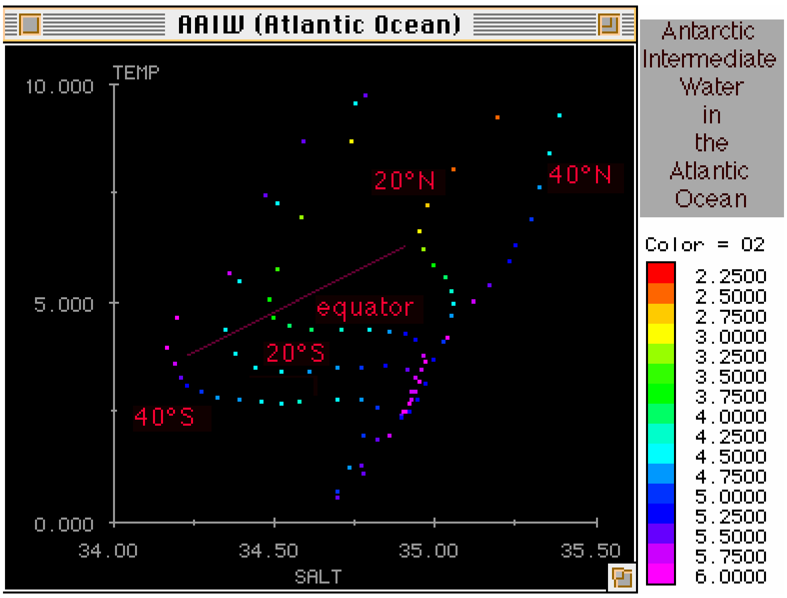


Figure 7.6. A series of TS-diagrams from the Atlantic Ocean, from 40°S to 40°N, showing the erosion of the salinity minimum associated with Antarctic Intermediate Water along the path of the water mass. The red line traces the salinity minimum produced by the relatively fresh Antarctic Intermediate Water.
Colour indicates oxygen content. Note how the oxygen content associated with the salinity minimum decreases from south to north, indicating the aging of the Antarctic Intermediate Water.

A complete description of water mass movement requires horizontal property distributions as well as vertical sections and TS-diagrams. It is then seen that the path of Antarctic Bottom Water in particular is strongly affected by the topography. For example, the deep basins of the eastern Atlantic Ocean are separated from the Southern Ocean by a sill and cannot be reached by Antarctic Bottom Water directly. They are filled through a gap in the Mid-Atlantic Ridge near the equator known as the Romanche Fracture Zone; in other words, flow of Antarctic Bottom Water in the eastern South Atlantic Ocean is *southward*, from the equator toward the pole. In the Pacific Ocean, input is mainly along 170°W (east of New Zealand), followed by spreading east and westward in the northern hemisphere; recirculation into the southern hemisphere occurs in the east. Input into the Indian Ocean is from the west, and in smaller quantities from the east.

Circulation in Mediterranean Seas

Mediterranean Seas are large bodies of water characterized by very restricted water exchange with the major ocean basins. This results in different hydodynamics and sets them apart from the remainder of the world ocean. While the circulation in most of the world ocean is dominated by wind-driven currents, the circulation in mediterranean seas is determined by thermohaline processes. Two basic types of circulation can be distinguished, the concentration basin and the dilution basin. Concentration basins occur where evaporation in the region exceeds precipitation; such mediterranean seas are therefore also sometimes called arid mediterranean seas. Examples are the (Eurafrican) Mediterranean Sea, the Red Sea and the Persian Gulf. Dilution basins occur when precipitation and river input exceed evaporation; such basins are therefore also known as humid mediterranean seas. Examples are the Black Sea, the Baltic Sea and the Australasian Mediterranean Sea (the seas of the Indonesian archipelago).

The circulation in mediterranean seas and their water exchange with the remainder of the world ocean differs strikingly between the two types (Figure 7.7).

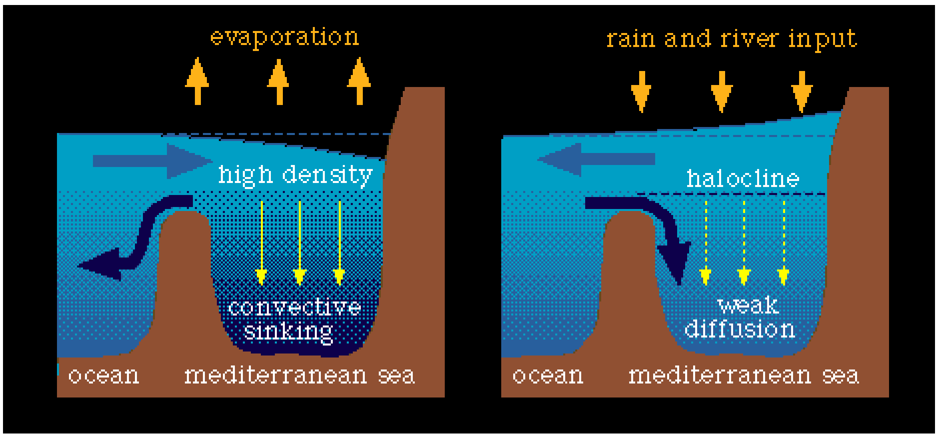


Figure 7.7. Circulation in mediterranean seas. left: in concentration basins; right: in dilution basins. In both situations water exchange across the sill occurs for two reasons, firstly to make up for the water loss or gain experienced by the mediterranean sea, and secondly as a result of the density and associated pressure difference between the mediterranean sea and the adjacent ocean. The second mechanism, which drives a flow from the region

of higher density to the region of lower density below a compensation flow in the opposite direction, produces by far the larger transport.

In concentration basins (Figure 7.8, Figure 7.9), evaporation increases the salinity of the surface waters, raising their density and producing convection. Deep water renewal is therefore a nearly continuous process, and the waters of the basin are well ventilated (have relatively high oxygen content) at all depth.

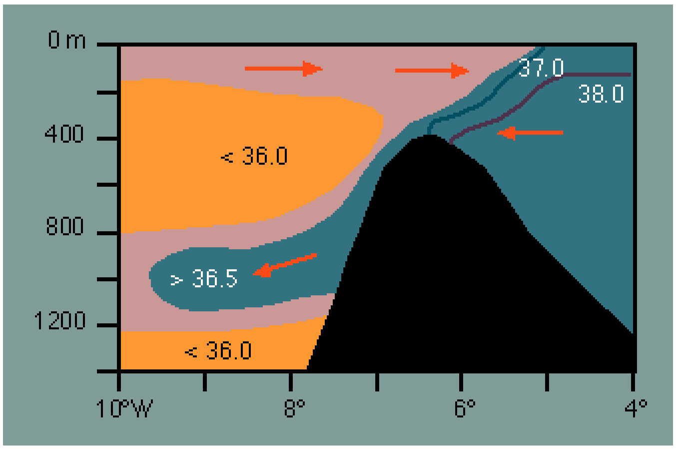


Figure 7.8. Water exchange between the Eurafrican Mediterranean Sea and the Atlantic Ocean proper across the sill in the Strait of Gibraltar as seen in the salinity distribution. The Mediterranean Sea is on the right, the Atlantic Ocean on the left. Numbers give salinity. On entering the Atlantic Ocean across the sill, the salty (dense) mediterranean water sinks to a depth of about 1000 m, where its density matches the density of the surrounding waters.

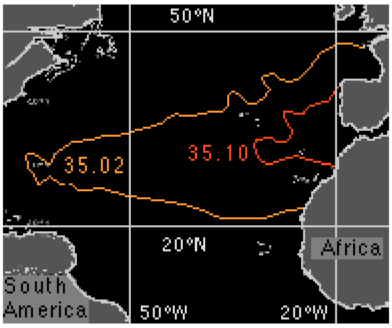


Figure 7.9. Spreading of Mediterranean Water in the Atlantic Ocean, indicated by the salinity at the salinity maximum produced by it. The depth of the spreading is typically just below 1000 m.

In dilution basins, the freshening of the surface waters resulting from excess rain and freshwater input from rivers reduces the density of the surface layer. This prevents the freshened water to reach the deeper layers. The result is the establishment of a fresh upper layer and a strong halocline. Water below the halocline is renewed only very slowly through mixing across the halocline and inflow of oceanic water through the connecting strait. As discussed in Lecture 5, oxygen at depth is consumed by remineralisation of nutrients. Oxygen content is therefore very low when active ventilation is inhibited through the stable halocline (Figure 7.10).

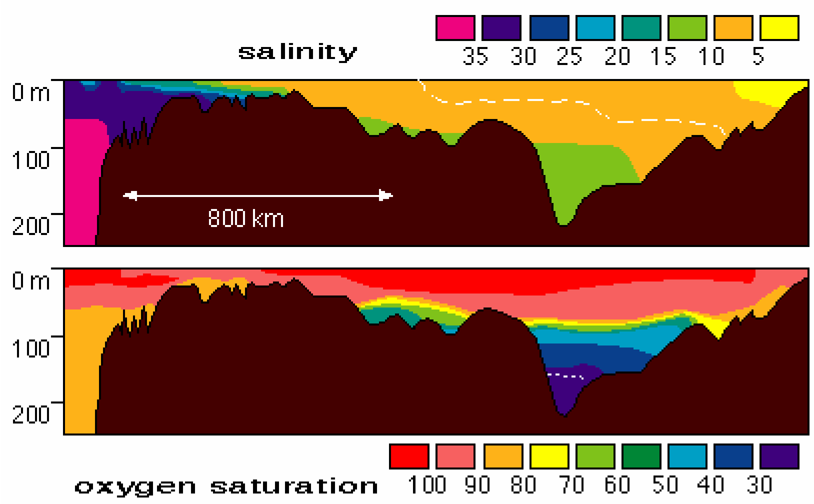


Figure 7.10. Salinity and oxygen (in % oxygen saturation) in the Baltic Sea. The North Sea is on the left, beyond the shallow sills of the Belt Sea between Sweden and Denmark where the depth is only 30 m or less. High salinities in the west indicate inflow of salty water from the North Sea. Note the rapid transition from salinities higher than 25 to salinities below 10 in the Belt region. Further east the Baltic Sea has very low salinity. (The broken line is the 7.5 isohaline.) Oxygen levels are close to or above saturation in the upper 50 m but get very low in the deeper layers. (The broken line is the 20% saturation contour.) Note the rapid decline of oxygen saturation near 100 m depth.

If the basin is large and the exchange with the open ocean very restricted, oxygen levels at depth can fall to zero, preventing the existence of higher marine life. Such conditions are occasionally found in some basins of the Baltic Sea. The Black Sea, which is more than 1500 m deep, is devoid of oxygen below 150 m depth.

Lecture 8

The ocean and climate

reviewed by: Alexandre Ganachaud and Scott B. Power

Ocean and atmosphere form a coupled system. The coupling occurs through exchange processes at the interface (the sea surface). These exchange processes determine the energy and mass budgets of the ocean. Quantities exchanged between the ocean and the atmosphere are:

In the energy budget:	radiative energy (including heat) momentum
In the mass budget:	freshwater, through - evaporation/condensation and - precipitation/river run-off minerals gases

Gases absorb radiative energy selectively. Some gases are transparent to the short wave radiation emitted by the sun but highly absorbent at the longer (infra red) wave lengths at which the earth emits radiative energy into space. High concentration of these gases in the atmosphere leads to a trapping of radiative energy in the atmosphere which manifests itself as an increase in the atmospheric temperature. These gases are therefore known as greenhouse gases. Carbon dioxide is a particularly important greenhouse gas.

The role of the ocean in the earth's climate and its capability of storing carbon dioxide (CO_2) is discussed in the following text, taken from The Ocean and Climate by R. W. Stewart (ims Newsletter 55/56 1991, Unesco, Paris). Some annotations, which are not part of the original text, were added to make the text more accessible to first year students; these annotations are in orange.

Begin of quotation

The ocean plays a role in the climate system which is complementary and of comparable importance to that of the atmosphere. It stores heat and releases it later, and often in a different place. It transports heat in amounts comparable with atmospheric transport. It both absorbs and releases carbon dioxide. (...) It is sometimes referred to as the "flywheel of the climate system" (...). Like a flywheel, the ocean stores energy, in this case thermal energy, when it is in large supply during the day or summer, and releases it when the energy supply is reduced or reversed during night or winter.

When it is heated the ocean responds by storing some of the heat and by increased evaporation. Because the heat is mixed down for some metres by the wind, temperature rises much less than it does on dry land under the same heating conditions. The evaporation has profound effects on the atmosphere and on climate. Water vapour released into the atmosphere importantly increases the greenhouse effect in the atmosphere. When it recondenses, the resulting heating of the air is a major source of energy for atmospheric motion.

When the ocean is cooled, it responds by generating vertical convective motions, which resupply heat to the surface. (This occurs because continuity of mass requires that cold water sinking from the surface is replaced by water from below. This water is - slightly - warmer than the sinking water and thus represents a supply of heat.) Thus the temperature fall is much less than over land under the same cooling conditions.

The overall result is that for the two thirds of the earth's surface covered by ice-free ocean, the temperature over the whole ocean ranges only from -2°C (the freezing point of salt water) to 30°C, and at any one place by hardly more than 1°C during the course of a day and 10°C during the course of a year. This range might be compared with that over dry mid continental areas, where the variation from place to place can be about 100°C, and during the course of the year in particular places about 80°C. Further, the relatively slow response of the ocean to heating and cooling results in the oceanic annual cycle being retarded relative to that in continental regions. (Much more energy is required to change the temperature of water than the temperature of air, so the ocean takes longer to heat up or cool down. As a result the ocean is still warming up in late summer when the air is still warmer then the water but already cooling, and it is still cooling down in late winter when the air is still cooler than the water but the atmosphere is already warming.) (...)

Such effects would be experienced even if the ocean were little more than a deep swamp. However, the ocean (...) moves. (...) In moving, it redistributes heat (and salt) in ways that are of central importance in determining the details of the earth's climate.

The North Atlantic provides a particularly notable example. In the tropical Atlantic, solar heating and excess evaporation over precipitation and runoff creates an upper layer of relatively warm, saline water. Some of this water flows north, through the passages between Iceland and Britain. On the way it gives up heat to the atmosphere, particularly in winter. Since winds at these latitudes are generally from the west, the heat is carried over Europe, producing the mild winters which are so characteristic of that region relative to others at similar latitudes.

So much heat is withdrawn (from the ocean and absorbed by the overlying atmosphere) that the temperature (of the water at the ocean surface) drops close to the freezing point. This water, now in the Greenland Sea, remains relatively saline, and the combination of low temperature and high salinity makes the water more dense than deeper water below it. Convection sets in and the water sinks - occasionally and locally right to the bottom. There it slides under, and mixes with, other water already close to the bottom. It spreads out and flows southward, deep and cold.

This thermohaline circulation: surface warm water flowing north, cooling, sinking and then flowing south provides an enormous northward heat flux. It amounts to 1PW (petawatt, 1 PW = 1 billion megawatts), fully comparable with that transported poleward by the atmosphere.

(...)

Water (from all depths) is in repeated contact with the surface and comes into approximate equilibrium with atmospheric concentrations of gases, including notably O_2, CO_2 and freons. The freons are inert (not influenced by biological processes or chemical reactions; their concentration is only affected by the mixing of water), and provide a valuable passive tracer for ocean movement. O_2 and CO_2, on the other hand, are strongly affected by biological activity. The surface layers of the ocean contain planktonic plants which, in the presence of sunlight, convert dissolved CO_2 into organic carbon. The plants are eaten by animals, which are in turn consumed by other organisms. Debris from these organisms falls out of the surface layers into the deeper water. On the way down, bacteria decompose some of the material, releasing CO_2 and absorbing O_2. As a result the deeper water is enriched in CO_2 and nutrients and depleted in O_2.

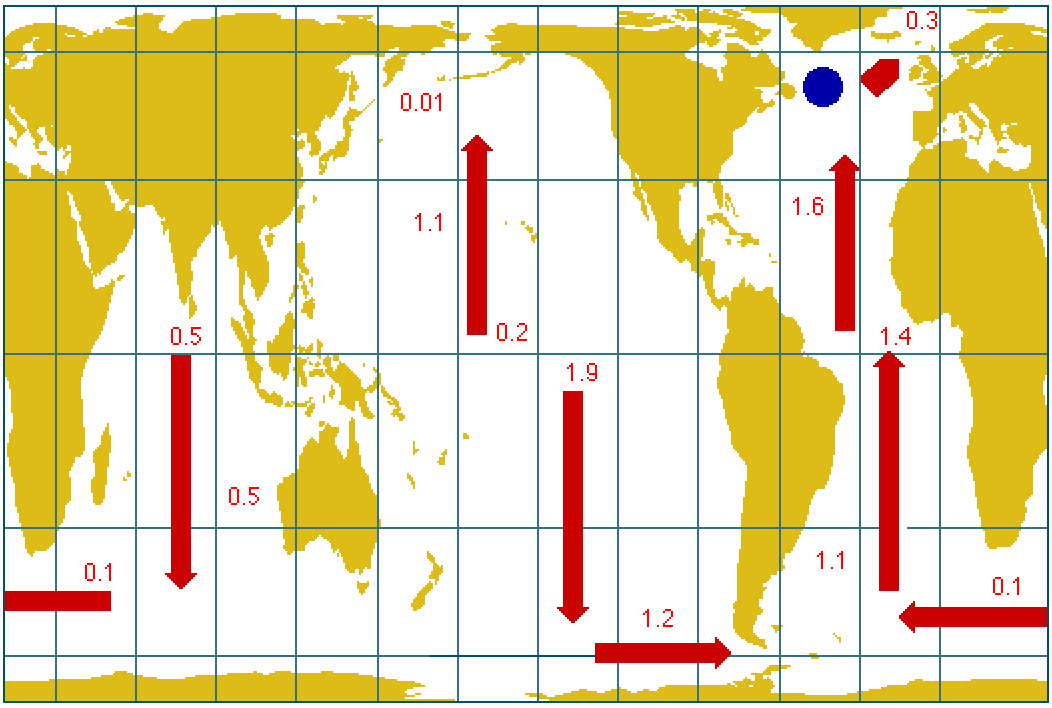
(...)

The ocean plays a key, but frequently understated, role in determining the earth's climate. Indeed any possibility of predicting the evolution of climate beyond a few weeks demands that ocean behaviour also be taken into account.

(...)

With respect to sensitivity to, and contribution to, long term climate change: there is every reason to believe that the ocean is now changing, in response to climate changes over the past few hundred years (the Little Ice Age). It can be expected to change further as anthropogenic influences (influences from human activity) become increasingly marked. The effect of the ocean on the atmosphere could be either to moderate or to intensify these changes. It will certainly modify them.

The map shows how the ocean circulation distributes heat throughout the world's oceans.



The above map shows how the ocean circulation distributes heat throughout the world's oceans (In petawatts (PW), 1 PW = one billion megawatts). Some essential ideas involved or demonstrated in this illustration (and not yet accepted in every detail by the scientific community) are:

1. the atmosphere is driven by heat given off by the ocean;
2. this heat is the major energy source for powering the strong westerlies in the North Atlantic (about 1 PW annually);
3. the concentration of heat flow from the ocean to the westerlies in the vicinity of the blue dot (off the coast of Newfoundland) can be very much greater during "peak loads" than the 100 megawatts per square km average for the region;
4. to supply that massive heat flow to the westerlies in the North Atlantic, the ocean circulation delivers heat from the Pacific and Indian Oceans;
5. the arrows show the direction of heat flows in the global ocean; and
6. the differences between the values shown at selected latitudes represent the heat (in PWs) contributed to the atmosphere from the ocean between those latitudes.

End of quotation

The thermohaline circulation described above ("surface warm water flowing north, cooling, sinking and then flowing south ") has become known as the Great Ocean Conveyor Belt (Figure

8.1): The water that sinks in the North Atlantic Ocean (North Atlantic Deep Water) enters the Antarctic Circumpolar Current and from there all ocean basins, where it rises slowly into the upper kilometre and returns to the North Atlantic in the permanent thermocline. Although this is only one of the circulation paths of North Atlantic Deep Water it is the most important one from the point of ocean/atmosphere coupling since it acts as a major sink for atmospheric greenhouse gases. The only other region of similar importance is the Southern Ocean where Antarctic Bottom Water sinks.

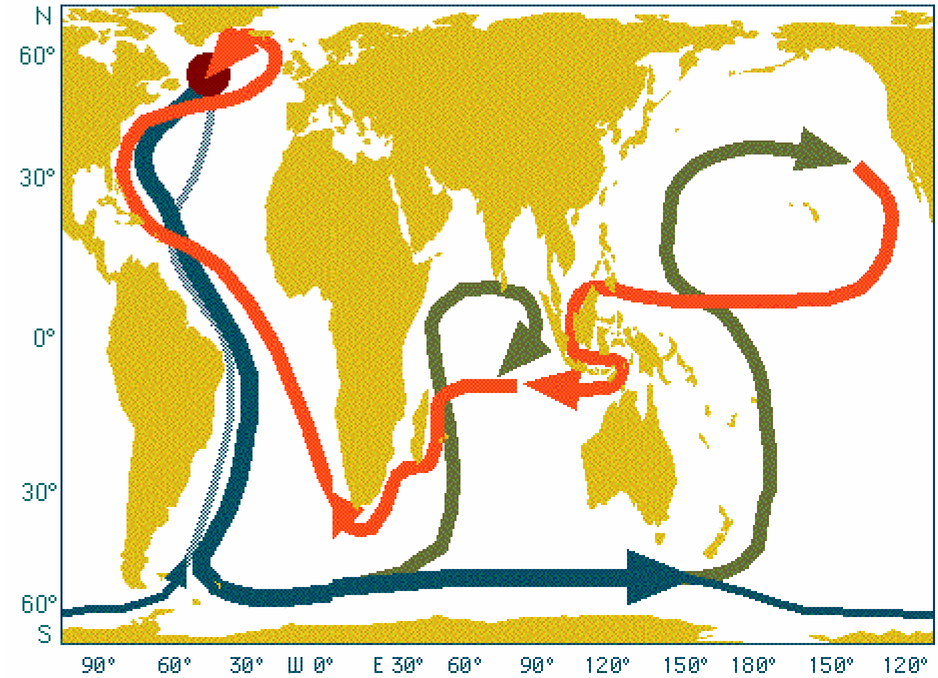


Figure 8.1. The path of North Atlantic Deep Water through the world ocean (The "Great Ocean Conveyor Belt"). Blue indicates deep currents, olive intermediate currents (about 1000 m depth) and orange surface currents.
Water sinks to the ocean floor in the North Atlantic Ocean (the brown circle). It moves southward as North Atlantic Deep Water and joins the Circumpolar Current. Some of it returns to the North Atlantic Ocean as a deep current, the remainder rises to intermediate depth, moves north across the equator, rises further into the upper ocean and returns to the North Atlantic Ocean with the surface currents.

Formation of North Atlantic Deep Water does not necessarily continue forever. The deep convection in the Greenland Sea occurs in a region where cold, fresh water and warm, saline water meet (Figure 8.2). Convection occurs when the warm, salty water gets cold enough to sink, just before its relatively high density forces it to slide underneath the fresh polar surface waters and continue as a subsurface current (the "Atlantic inflow" in Figure 8.2). Convection can be inhibited by a number of processes. If the climate gets warmer, additional melting of ice will increase the volume of the Arctic outflow of cold, fresh water and push the region where the warm, salty current is forced to underneath the fresh polar surface waters further to the south. The warm, salty water is then insulated from atmospheric cooling and will no longer sink. This will stop the conveyor belt. As a result, Europe will become much colder, more ice will form in the Arctic, outflow of cold, fresh water will be reduced, and the conveyor belt will be active again. It is seen that the ocean can support two alternate circulation systems as the two states of an oscillation system. There is geological evidence that the conveyor belt is inactive during ice ages.

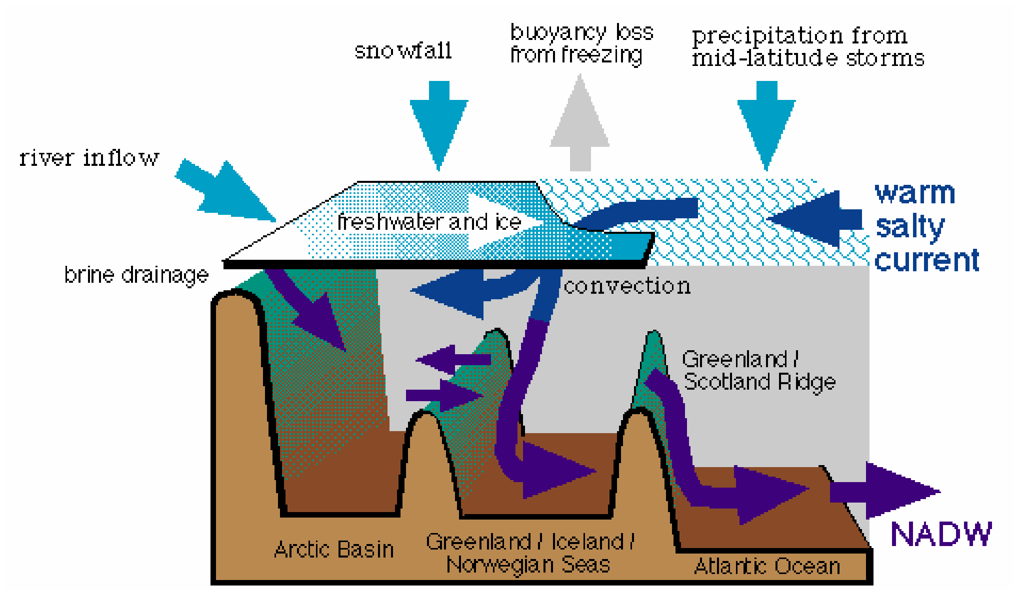


Figure 8.2. A sketch of the processes which lead to the sinking of water in the Greenland Sea. NADW = North Atlantic Deep Water.
The water from the warm and salty Gulf Stream Extension feeds the Atlantic inflow into the Arctic Basin and the convection in the Greenland Sea. The convection can be inhibited by a southward advance of freshwater and ice export from the Arctic Ocean. See the text for a more detailed discussion.

The question whether and to which degree the thermohaline circulation is susceptible to human activity is the subject of intense research in numerous institutions around the world.

El Niño and the Southern Oscillation (ENSO)

The discussion of ocean circulation changes and ice ages above gave one example of oscillatory behaviour of the coupled ocean/atmosphere system. Another example, on a time scale short enough to be experienced during a human life span, occurs in the Pacific Ocean and is known as ENSO, which stands for El Niño - Southern Oscillation. The Southern Oscillation is the term for a large-scale oscillation of air pressure observed in the tropics around the globe and particularly clearly over the tropical Pacific Ocean, where air pressure is high in Darwin when it is low in Tahiti (or the central and eastern Pacific Ocean in general) and vice versa. Figure 8.3 shows the effect of the Southern Oscillation on air pressure and rainfall. It is seen that high air pressure at Darwin is linked with high rainfall in the central Pacific.

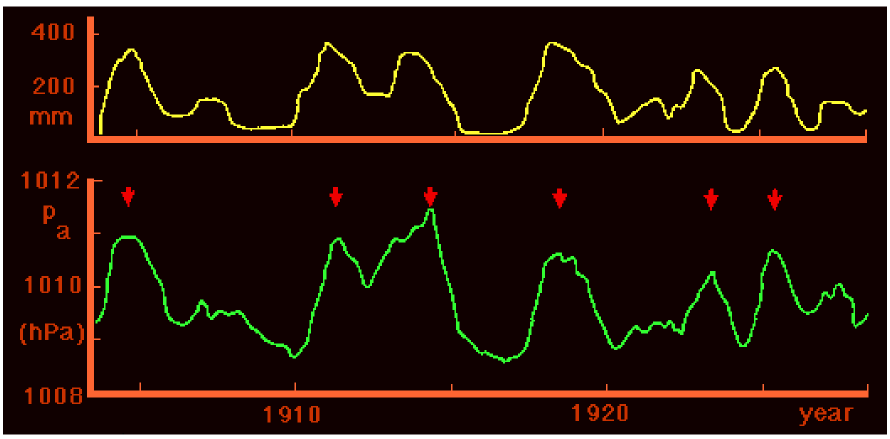


Figure 8.3. Darwin air pressure (bottom) and rainfall at central Pacific islands (top). Air pressure at Tahiti is the mirror image of air pressure at Darwin. Occurrences of high Darwin air pressure and high central Pacific rainfall

(arrows) are called ENSO events. From Tomczak and Godfrey, *Regional Oceanography: an Introduction*, Pergamon, New York (1994), 422 pp.

El Niño is the name for the oceanographic side of the phenomenon. One of the richest fishing regions of the World Ocean, the South Pacific coastal upwelling region along the coast of Peru, Chile, and Ecuador, occasionally experiences an influx of nutrient-poor, warm tropical water which suppresses the upwelling of nutrients. The *anchoveta* which inhabit these waters in their millions forming the nutritional basis for a huge bird population and the stock for an important fish meal industry, depend on the supply of nutrients to the surface layer. They avoid the warm nutrient-poor water, which causes mass mortality amongst the birds. If the extent of the tropical influx is very severe, mass mortality can occur among the fish as well; hydrogen sulphide from decaying fish has been known to blacken the paint on ships in Callao harbour. When this occurs it occurs usually just before Christmas - thus the name "El Niño" (the child) which relates the event to the birth of Christ. The high temperatures along the South American coast last for about a year or more before conditions return to those which prevailed before the influx of tropical water.

A simplified description of the mechanism how ocean and atmosphere interact to cause an ENSO events starts from the effect of sea surface temperature on winds. Figure 8.4 indicates that the two convergence zones in the atmosphere coincide with regions of high sea surface temperatures. This is because the air is heated where the water is warm; it rises, producing a convergence of the winds above the sea surface (seen in high average cloud cover, Figure 8.5) - in other words, sea surface winds blow towards regions of high sea surface temperature. This results in accumulation of warm water, which increases the heating; the air rises faster, wind speed increases - a positive feedback is established.

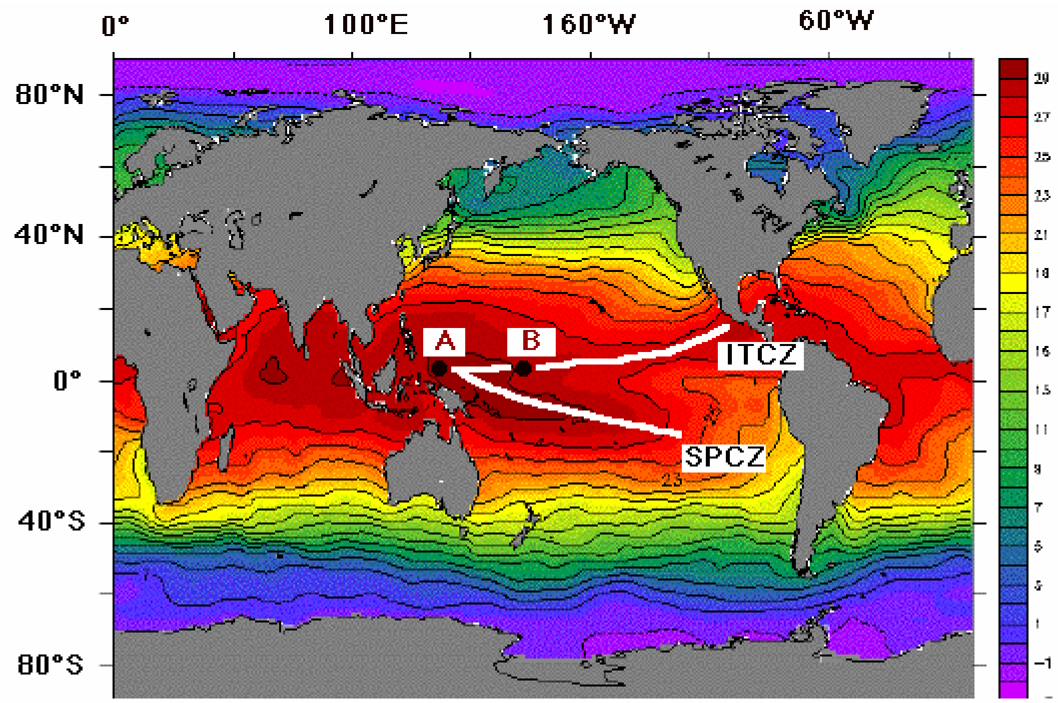


Figure 8.4. The Intertropical Convergence Zone (ITCZ) and South Pacific Convergence Zone (SPCZ) in relation to average sea surface temperature. See the lecture text for an explanation of points *A* and *B*.

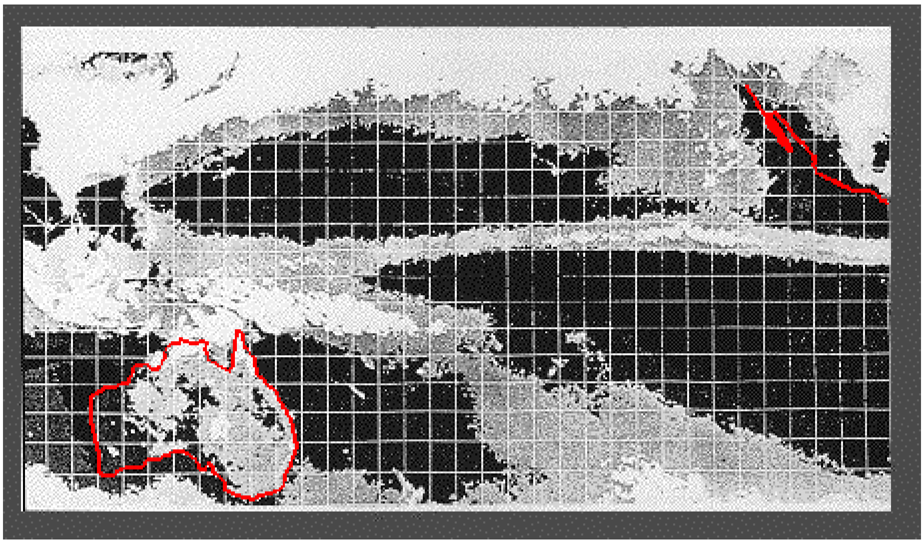


Figure 8.5. The Intertropical Convergence Zone (ITCZ) and the South Pacific Convergence Zone (SPCZ) as seen in satellite cloud images. The figure is a composite of many months of observations, which makes the cloud bands come out more clearly. High cloud density indicates condensation from rising air and thus high rainfall. The figure covers the region 40°S - 40°N, 97°E - 87°W; the grid gives every 5 degrees latitude and longitude (Australia is in the lower left, parts of North America with Baja California in the upper right). From Tomczak and Godfrey, *Regional Oceanography: an Introduction*, Pergamon, New York (1994), 422 pp.

Suppose now that through some disturbance the region of highest temperature is shifted from the region where the ITCZ and the SPCZ meet (point A in Figure 8.4) to a point somewhere further east (point B). Winds continue to blow towards the highest sea surface temperature; so the winds to the west of that point will reverse their direction and change from Trades to westerlies. Again, this pattern will be reinforced through positive feedback. The centre of rainfall is shifted from A to B; drought conditions are observed in Australia. This is often accompanied by tropical cyclone development (Figure 8.6).

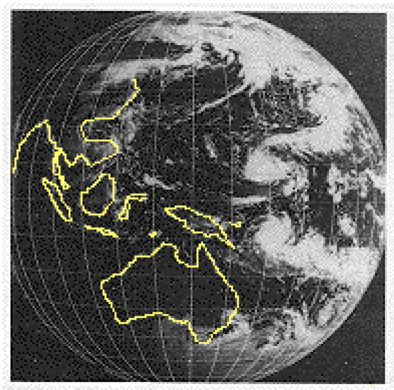


Figure 8.6. Cloud cover over the Pacific Ocean, showing a tropical cyclone pair in formation produced by wind shear between the equatorial westerlies of an ENSO event and the Trades.

The change from one state of the ocean/atmosphere feedback system to the other requires unstable conditions in the atmosphere. Such conditions occur usually around May or June when the atmospheric circulation over the adjacent Indian Ocean changes from Northeast to Southwest Monsoon. Whether an ENSO event occurs in a particular year is therefore usually decided in May.

The westerly winds in the western equatorial Pacific trigger a solitary internal wave of large scale (several hundreds of kilometers in length and about 400 km across) which travels eastward along the equator, advecting warm tropical water into the South American coastal upwelling region. The detailed dynamics are complicated and involve several types of long oceanic waves of very low frequency which take between 1 - 4 months to cross the equatorial Pacific Ocean and alter the temperature of the upper ocean thousands of miles from where they were generated. What is important in the process is that the original disturbance in the ocean/atmosphere system which occurred in the western Pacific Ocean early in the year produces a suppression of coastal upwelling and an influx of tropical water along the coast of Peru, Chile and Ecuador later in the year. The suppression of upwelling is most intense in November - December and does not disappear until well into the following year.

To conclude this lecture we look at the ENSO event of 1997/98. This was a particularly strong and long lasting ENSO event, not necessarily typical of others; but all ENSO event are different, and the 1997/98 El Niño was observed by satellite, so we have a particularly good data base.

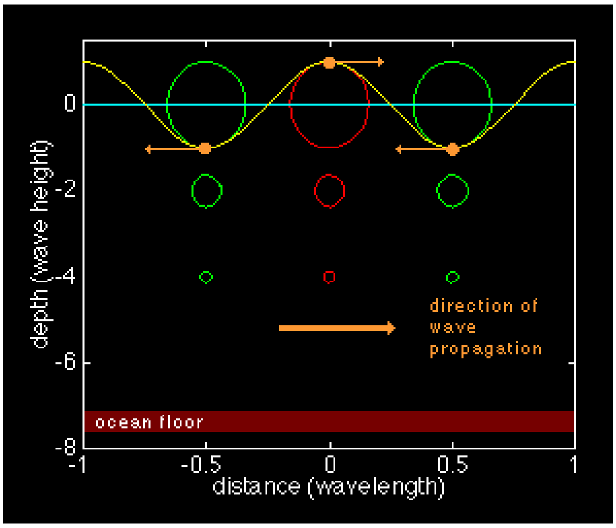
Lecture 9

Waves

reviewed by: Ivan Lebedev and Yin Soong

Waves are periodic deformations of an interface. Surface waves in oceanography are deformations of the sea surface, ie the atmosphere-ocean interface. The deformations propagate with the **wave speed**, while the particles describe orbital or oscillatory motions at **particle speed** and remain at the same position on average.

In deep water, particle paths are circles. In shallow water, the particle paths flatten to ellipses (Figure 9.1).



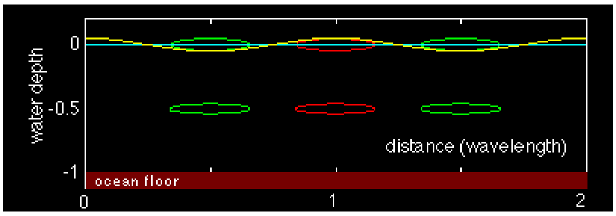


Figure 9.1. Particle movement in deep and shallow water waves. The diagrams use vertical exaggeration for clarity. First diagram: deep water waves. Second diagram: shallow water waves.
In deep water waves particles move on circles, in shallow water waves particles move on very flat ellipses. Particle movement decreases rapidly (exponentially) with depth in deep water waves but remains essentially the same over the entire water depth in shallow water waves.
In both cases, particles under wave crests move in the direction of wave propagation; particles under wave troughs move against the direction of wave propagation.

The change from deep to shallow water waves is observed when the wavelength λ becomes larger than twice the water depth h. A change in wave properties occurs also at $\lambda = 20h$. It is therefore useful to distinguish between

```
|________________________________|________________________________|_____________________________
0            < λ <               2h           < λ <              20h        < λ

   deep water waves                 transitional waves                shallow water waves
    or short waves                                                       or long waves
```

It is most important to note that the distinction between deep and shallow water waves has little to do with absolute water depth but is determined by the *ratio* of water depth to wave length. The deep ocean can be shallow with respect to waves provided the wave length exceeds twice the ocean depth. This is the case for example with tides.

Wave Classification

Ocean waves can be classified in various ways. One classification uses the forces which generate the waves. In ascending order of wave lengths we have:

1. Meteorological forcing (wind, air pressure); sea and swell belong to this category.
2. Earthquakes; they generate tsunamis, which are shallow water or long waves.
3. Tides (astronomical forcing); they are always shallow water or long waves.

Another classification is based on the frequency spectrum representation of all oceanic waves and distinguishes between capillary waves, gravity waves, long period waves, tides and transtidal waves (Figure 9.2). Yet another classification is based on the restoring forces responsible for returning the water particles to their average position in the water column (Figure 9.2).

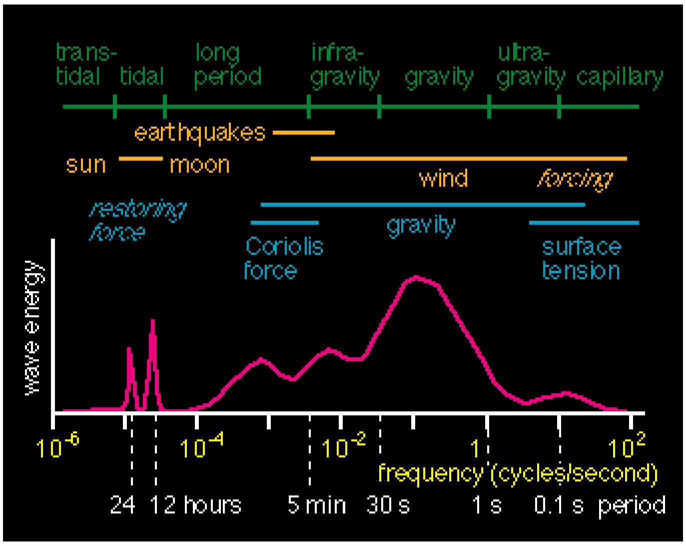


Figure 9.2. Sketch of the relative amounts of energy as a function of wave frequency in ocean waves. The top line (green) gives the classification based on period, the line below (gold) the classification based on the wave-generating force, and the bottom line (blue) the classification based on the restoring force.

Description of Waves

The simplest way of looking at waves is the concept of a wave as a harmonic oscillation (Figure 9.3).

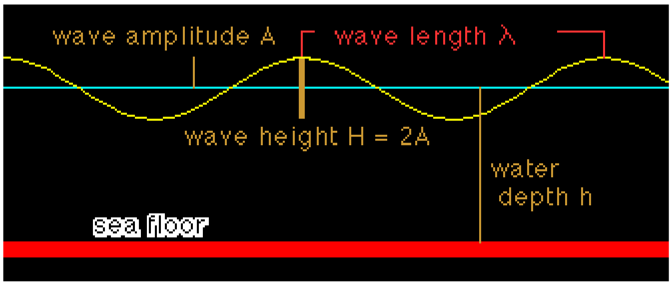


Figure 9.3. Sketch of a harmonic or ideal (sinosoidal) wave. H is exaggerated against λ for clarity. The narrow vertical bar gives the wave amplitude A (the distance between mean water level and wave crest); the heavy vertical bar gives the wave hight H (the difference between wave trough and wave crest, or twice the wave amplitude).

It can then be described by its

- period τ
- frequency $f = 1 / \tau$
- angular frequency $\omega = 2 \pi / \tau$
- wavelength λ
- wave speed $c = \lambda / \tau$
- wave height $H = 2A$ (A = amplitude)
- wave steepness $\Delta = H / \lambda$

Superposition of two waves with nearly equal frequencies ω_1 and ω_2 produces wave groups or packets (Figure 9.4).

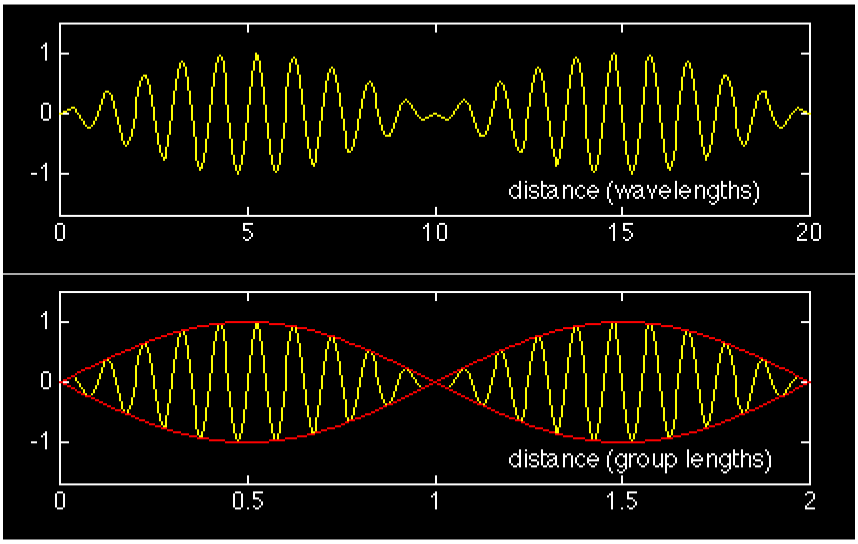


Figure 9.4. Wave groups produced by two harmonic waves of nearly identical frequencies. The upper diagram shows the waves as they are seen by an observer. The lower diagram shows the same waves but includes the envelope of the wave groups or packets as a red line.
The waves (yellow lines) travel at wave speed c; the wave groups or packets (red lines) travel at group velocity c_g. The packets transport the energy contained in the wave field, which means that the energy also travels at group velocity c_g.

Individual wave crests travel with phase velocity (identical to the wave speed) c; wave packets travel with group velocity

$$c_g = c - \lambda\, dc/d\lambda$$

(Note for the mathematically inclined: An equivalent but more conventional expression often used in oceanography is $c_g = d\omega/dk$, where k is the wave number.)

Although wave particles remain at the same position on average, waves transport energy in packets. Energy propagates with group velocity; it can move faster or slower than individual wave crests. This is called dispersion.

Normal dispersion

In normal dispersion, c increases with λ, ie the crests of long waves travel faster than the crests of short waves. As a consequence $c_g < c$, ie energy travels slower than the wave crests. This occurs with gravity waves such as swell.

Nondispersive waves

Here, $c_g = c$, ie all wave crests travel at the same speed, and energy propagates at the same speed.

Anomalous dispersion

This is found when $c_g > c$. Capillary waves are an example. Energy propagates faster than wave crests, and short waves travel faster than long waves.

In most oceanic situations the wave steepness is very small and the wave speed is then given by

$$c = \sqrt{\frac{g\,\lambda}{2\pi}\tanh\frac{2\pi\,h}{\lambda}}$$

(valid for $\Delta << 1$, or $\lambda >> H$)

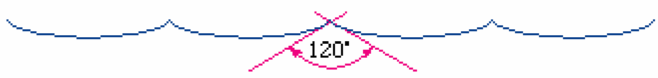
Here, H is the wave height and h the water depth. The formula can be simplified depending on the ratio of λ against h:

Deep water wave speed (short waves; depth is larger than 1/2 of the wavelength)	$c = \sqrt{\frac{g\,\lambda}{2\pi}}$	$c_g = c / 2$ (normal dispersion)
Shallow water wave speed (long waves; depth is less than 1/20 of the wavelength)	$c = \sqrt{g\,h}$	$c_g = c$ (nondispersive waves)

As an example, consider waves produced by a distant storm. In the open ocean such waves travel as deep water waves; their wave speed therefore depends on their wavelength λ. Thus long waves travel fastest and arrive at distant locations first. They are experienced as swell. When swell approaches a beach the water depth decreases and there comes a point when the waves change from deep water waves to shallow water waves. As a consequence the wave speed *c* decreases as the depth decreases and the waves bend inwards; the wave front becomes progressively more parallel to the beach.

Waves of finite amplitude

The harmonic oscillation concept gives a good description of waves of very small steepness. On the approach of a beach, or during the period of active formation by wind, wave steepness is not small enough and the wave profile deviates from the harmonic profile. As the steepness increases, the wave profile becomes cusped:



Eventually, the waves break. The limiting values before breaking are 120° for the cusp angle and a steepness $\Delta = 1/8$. Cusped waves do not have closed particle paths but are associated with a net transport of water (not just energy). This transport is called Stokes drift, after its investigator.

Short waves (deep water waves)

Short waves in the ocean are wind generated waves. They are divided into sea and swell. Sea includes all waves generated by the local winds, while swell refers to waves generated by distant wind fields.

The effect of the wind on the state of the sea depends on the distance over which the wind can blow without hindrance before it reaches the observation point. This distance is known as the fetch. The presence of a coast limits the fetch for winds blowing from land to sea. In the open ocean the fetch is usually determined by the size of the weather system that produces the wind.

Another wave determining factor is the time during which the wind blows unabated with a given strength. At any given wind speed it takes a certain time for the waves to build up to a steady state. The time required to reach that state (where the waves do not grow any further) is known as the duration of the wind.

At any given instant in time the sea state is never a single harmonic oscillation, but appears to the observer as a chaotic process. Therefore ways have to be found to describe the wave conditions in terms of measurable statistical quantities. Two approaches are used:
1) Determination of significant wave parameters (description in the time domain)
2) Determination of the wave spectrum (description in the frequency domain)

The combination of many measurements has resulted in wave parameter estimates for fully developed seas (Figure 9.5). A fully developed sea is one for which the fetch and duration are not limiting ie no further growth occurs, as momentum and energy loss through breaking balances input from the wind.

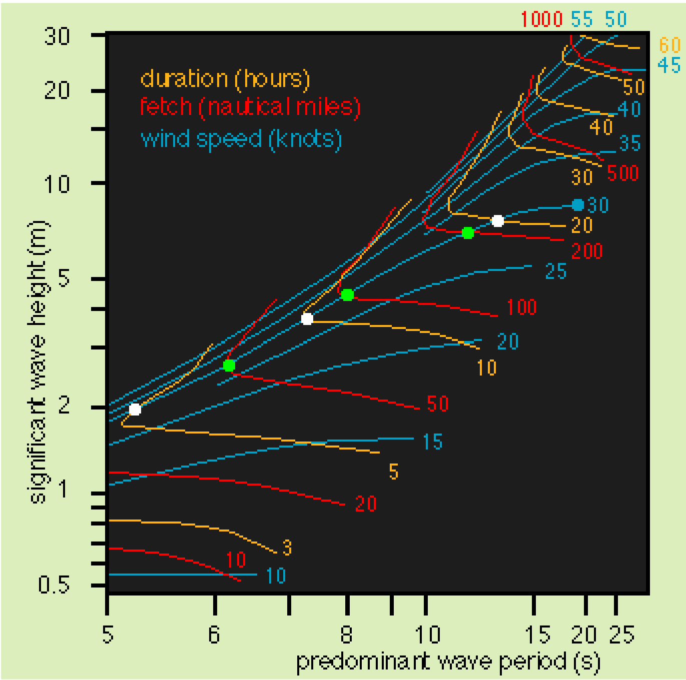


Figure 9.5. Wave height and period as functions of wind speed, duration and fetch. The diagram allows the estimation of wave build-up to a fully developed sea. The white and green dots give two examples:
1.) The series of white dots shows, from left to right: A wind of 30 knots produces waves of 2 m wave height with a period of 5.2 seconds after it has blown for 5 hours over a minimum fetch of about 35 nautical miles. After 10 hours the same wind will have produced waves of 3.6 m wave height with a period of 7.2 seconds, provided it blew over a minimum fetch of about 80 nautical miles. After 20 hours waves produced by the same wind will have reached a height of 7 m, provided the wind blew over a minimum fetch of about 250 nautical miles.
2.) Alternatively, the series of green dots shows, from left to right: A wind of 30 knots that blew over a distance of 50 nautical miles produdes waves of 2.6 m wave height with a period of 6.1 seconds, provided it blew for at least 6.5 hours. If it blew over a distance of 100 nautical miles the generated waves are 4.2 m in wave height and 8 seconds in period, provided the wind blew for at least 11 hours. Over a fetch of 200 nautical miles the same wind generates waves of 7 m wave height and a period of 12 seconds, provided it blew for at least 18 hours.
The fully developed sea from a 30 knot wind is reached at the blue dot (the end of the blue line), when the wind blew for at least 22 hours over a fetch of at least 350 nautical miles. It is characterized by waves of 8 m wave height with a period of 18 seconds.

(1 knot = 1 nautical mile per hour or approximately 1.8 km/hr)

The graph shown in Figure 9.5 applies to situations where waves are generated by local wind. Figure 9.6 shows wave properties for situations where the waves are generated remotely. It assumes fully developed sea conditions in the generation region.

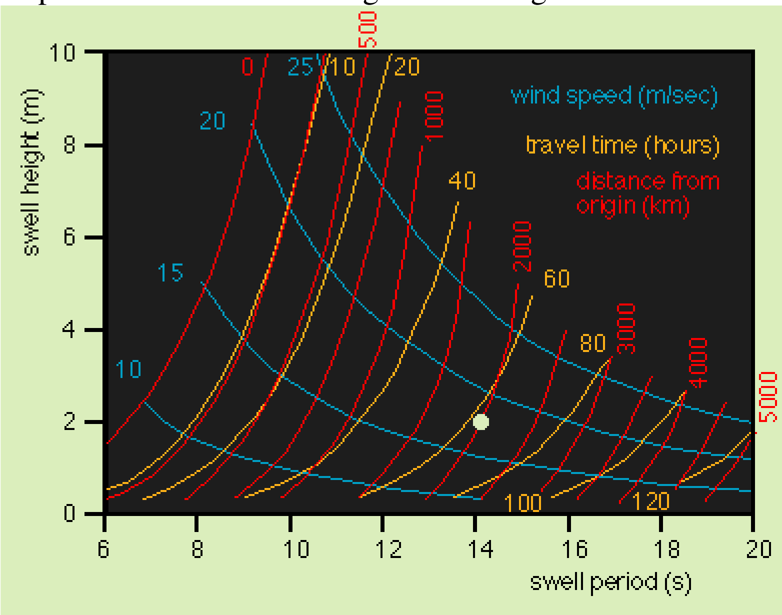


Figure 9.6. Height and period of swell as functions of distance from the generation region, wind speed in the generation region and travel time to the observation point. Example: Swell of 2 m height and 14 second period (the dot in the diagram) is produced by a wind blowing at 18 m s^{-1} at 2000 km distance 62 hours ago.

For many marine applications, for example the routing of ships or the design of platforms, only the heighest waves are of interest. The quantity **significant wave height** has therefore been introduced. It is defined as either $H_{1/3}$ or $H_{1/10}$, ie as the average of the 1/3 or 1/10 heighest waves over an observation period. (The use of $H_{1/3}$ is more common than the use of $H_{1/10}$). From observations, the largest wave height H_{max} is related to the significant wave height by

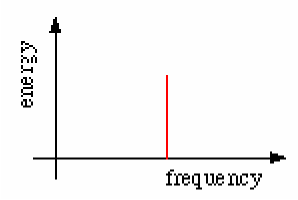
$$\frac{H_{max}}{H_{1/3}} = 1.45 \qquad \text{and} \qquad \frac{H_{max}}{H_{1/10}} = 1.3$$

Measured maximum wave heights depend to some degree on the length of the observation period; different values for H_{max} are found from a 10 minute time series compared to a 3 hour time series.

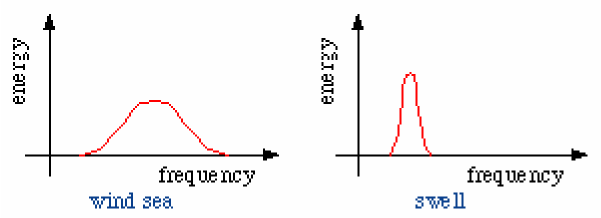
Statistical description of waves

The relations between largest wave height and significant wave height indicate that the sea state has certain statistical properties. A statistical description is based on the representation of the wave field through the energy spectrum. For a given frequency, wave energy is proportional to the square of the amplitude.

An energy spectrum shows wave energy as a function of wave frequency. A single harmonic wave has a 'monochromatic' spectrum:



Statistical description assumes that waves of all frequencies and corresponding wavelengths are present. It does not attempt to describe the form of the sea surface but concentrates on wave energy. In a wind sea with random distribution of wave energy over all wave frequencies, the theoretical form of the energy spectrum is that of a Gaussian or 'normal' distribution. Where only swell is present the energy is concentrated near the swell frequency and the spectrum is much narrower:



In real oceanic situations energy is not randomly distributed but cascades from the shortest possible waves stirred up by the wind to the longer wavelengths. As a result the shape of ocean wave spectra depends strongly on the wind speed. Figure 9.7 shows observed energy spectra for fully developed seas at various wind speeds. Note that the spectrum has a normal distribution only for very low wind speeds; as the wind speed increases, waves of short period are still present but most of the energy is found in longer period waves. The spectrum then drops off rapidly at longer periods; it is "skewed" towards the longer periods.

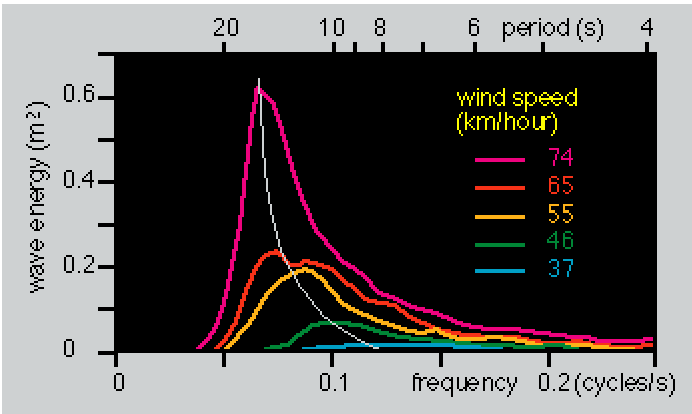


Figure 9.7. Wave energy as a function of wave frequency for fully developed seas under different wind speeds. The colour of the spectra indicates wind speed in km/h. The thin line shows the shift of the dominant wave period (the waves that contain most of the energy) towards longer periods as the wind speed increases.

Final wave decay occurs during breaking on a beach, which occurs when the particle velocity becomes larger than the phase velocity (wave speed). In this state the waves transport both energy and mass towards the beach. While much of the energy is dissipated in mechanical reworking of the beach, the mass moved by the waves has to be returned to the sea. This occurs at regular intervals along the beach in so-called rip currents, strong seaward movement of the water along the sea floor ("undertow").

Lecture 10

Long Waves (shallow water waves)

reviewed by: Yin Soong

Short waves (deep water waves) show normal dispersion, i.e. wave speed depends on the period, with the longer period waves moving faster than the shorter period waves (and the longer period waves have the longer wave lengths).

In contrast, long waves (shallow water waves) are non-dispersive: Their wave speed is independent of their period. It depends only on the water depth, in the form

$$c = \sqrt{g\,h}$$

(c is wave speed, h water depth, g gravity).

The velocity structure in a long wave is described by

$$u = \frac{g\zeta}{\sqrt{g\,h}}$$

where ζ is the time dependent surface elevation (wave amplitude) and u the horizontal particle velocity. It follows that u is independent of depth and the vertical particle velocity varies linearly with depth. Particles move on very flat elliptic paths in nearly horizontal motion.

Tsunamis

Tsunamis are long waves generated by submarine earthquakes. Tsunami is Japanese for "harbour wave". They are often called tidal waves; but this is a misnomer, since tsunamis have nothing to do with tides.

Before 2004 the strongest tsunami in known history was produced by the eruption of the Krakatau of the Sunda Island group in 1883. It reached a wave height of 35 m and claimed 36,830 lives. Four tsunamis with heights in excess of 30 m have been documented in the Pacific Ocean since 684 A.D. A strong tsunami in the Atlantic Ocean was observed in 1755 after an earthquake near Lisbon (Portugal).

In the vicinity of the epicentre of an earthquake, tsunamis can result in extreme wave heights. Once they reach the open ocean and travel through deep water tsunamis have extremely small amplitudes but travel fast, in 4000 m water depth at about 700 km/h. (This speed can be estimated by using the wave speed equation given above: We have g = 9.8 m s^{-1}, h = 4000 m, so $(9.8 \times 4000)^{1/2}$= 200 ms^{-1}= 700 km/h.) On approaching a coast they build up wave height again through shoaling. The period of tsunamis is in the range 10-60 minutes. Figure 10.1 shows a record of a tsunami from an Alaskan earthquake recorded in Hawaii.

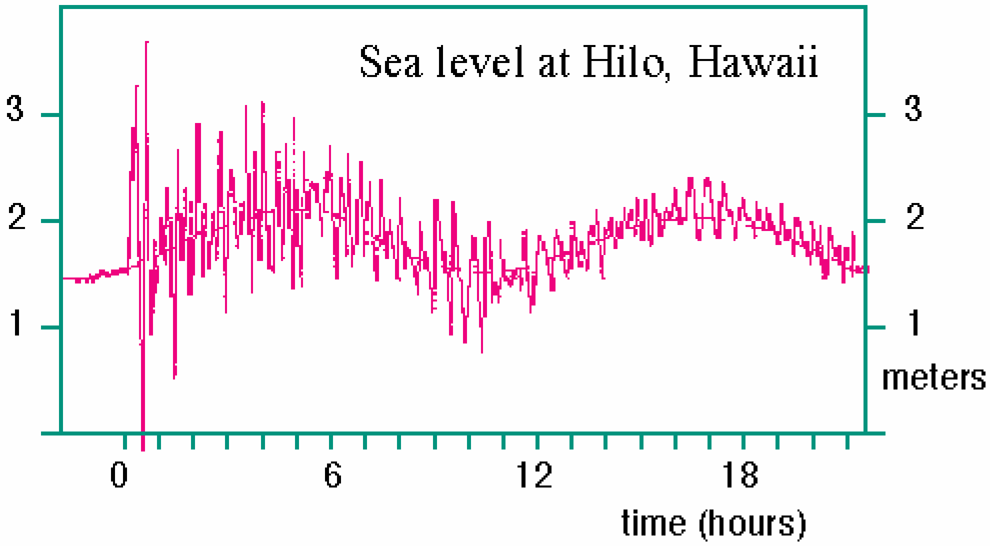


Figure 10.1. Passage of a tsunami as seen in a sea level record from Hilo, Hawaii. The slowly varying broken line indicates the rise and fall of the tide as it would have been observed without the tsunami. The observed sea level shows high frequency variations with a period of approximately 20 minutes and an initial amplitude of nearly two meters (total tsunami wave height 3.7 m)

Tsunamis were used to estimate the depth of the ocean in 1856, when direct depth measurements were virtually impossible, by observing their phase speed. The result for the North Pacific was 4200 - 4500 m, which was a considerable improvement on the previous estimate of 18,000 m.

The most destructive tsunami known occurred on 26 December 2004. It was generated by an earthquake in the vicinity of the Andaman Islands and northern Sumatra and caused death and destruction in countries around the Indian Ocean. The death toll is estimated at between 265,000 and 320,000, although a final accurate figure may never be known.

Because of the destructive force of tsunamis, a tsunami warning system has been set up. It uses seismographic observations of earthquakes and calculates arrival times around the coastlines of the oceanic basin. Another possibility is the monitoring of compression waves linked with volcanic eruptions; they travel at the speed of sound (1500 m s^{-1}) in the SOFAR channel. No warning system is available for areas in the vicinity of the epicentre.

Figure 10.1a shows the passage of the tsunami of 26 December 2004 through the Seychelles.

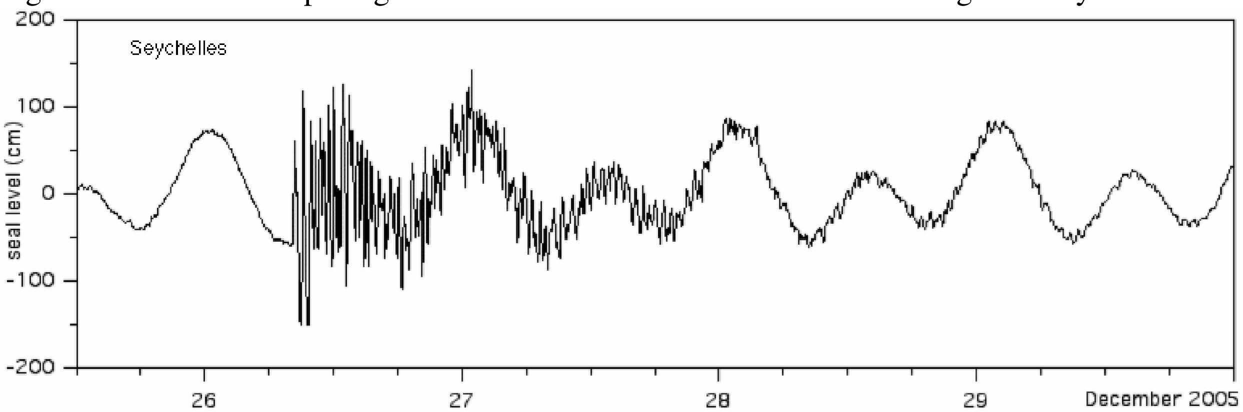


Figure 10.1a. Passage of the tsunami of 26 December 2004 as seen in a sea level record from the Seychelles. Data from the Seychelles Meteorological Office.

Seiches

Seiches are standing waves in closed or semi-enclosed basins. Consider a basin of length *L* and depth *h* with long wave speed

$$c = \sqrt{g\,h}$$

The time it takes a wave to travel through distance *L* is

$$\frac{L}{\sqrt{g\,h}}$$

Reflection occurs at the wall and the same time is needed to return to the starting point. Thus the basic period of a standing wave in the basin is

$$T_1 = \frac{2\,L}{\sqrt{g\,h}}$$

This is the period for the free oscillation of lowest (first) order. Higher order waves are possible with periods T_1/n for order *n*. The order is given by the number of nodes in the surface oscillation. Figure 10.2 shows the first order seiche (of period T_1), Figure 10.3 the second order seiche (of period $T_2 = 1/2\ T_1$).

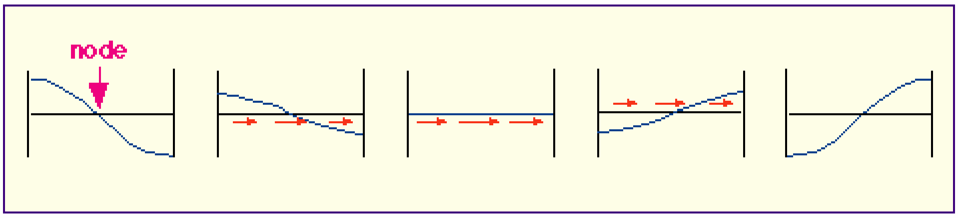


Figure 10.2. Sketch of a first order seiche at time intervals of 1/8 T_1 (ie over 1/2 of its period). Arrows indicate the direction of water movement. Water movement is in the opposite direction during the other half of the wave's period.

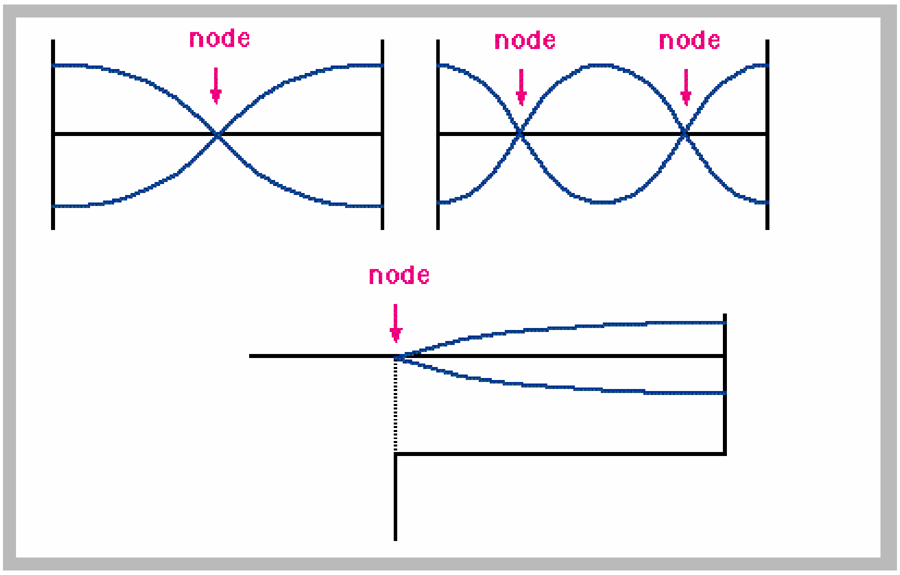


Figure 10.3. Sketch of seiches of different order, showing the position of the sea surface at the two opposite phases of the wave (1/2 period apart). The first diagram shows a first order seiche; it is identical to the seiche shown inFigure 10.2. The second diagram shows a second order seiche. The third diagram shows a first order seiche in a

bay which is open to the sea; here, the node is at the mouth of the bay and the wavelength is twice the wavelength observed in a closed basin.

Figure 10.4 shows a first order seiche in the Baltic Sea, where such oscillations of water level are usually produced by storms systems travelling across. The storm triggers a free oscillation (seiche) which then continues for several days before it is damped by bottom friction.

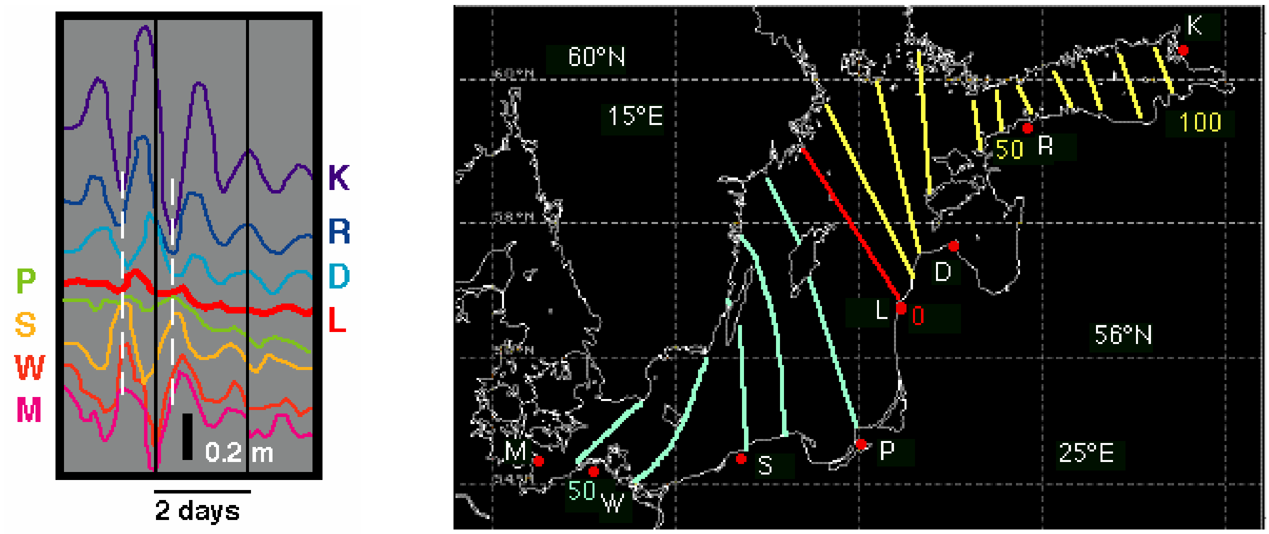


Figure 10.4. Observations of a first order seiche in the Baltic Sea.
First diagram: Observed water level variations at Koivisto (K), Reval (R), Domesnaes (D), Libau (L), Pillau (P), Stolpmünde (S), Warnemünde (W) and Marienleuchte (M). The broken white lines indicate the period of the seiche (slightly more than 1 day). The black bar gives the vertical scale.
Second diagram: Observed wave amplitude. Contour interval is 10 cm. Note the node at Libau (indicated by the red zero contour). At locations in the yellow contoured region the seiche is in the opposite phase to locations in the aqua-green contoured regions. The same phase relationship can be seen in the left diagram by comparing, for example, the records of Koivisto and Reval with the record of Stolpmünde.

If the basin is open, the connecting line to the open sea has to be a node (Figure 10.3). The corresponding period for the lowest order wave is therefore twice that of the lowest order seiche that would exist if the basin were closed (The effective wave length is twice the length of the basin). It is determined in analogy to the frequency determination of organ pipes and is

$$T_1 = \frac{4L}{\sqrt{g h}}$$

Higher order seiches, with period T_1/n, are again possible.

Internal Waves

It was said before that waves are periodic movements of interfaces. If the water column consists of an upper layer and a denser lower layer, the interface between the layers can undergo wave motion. This motion, which does not affect the surface and mostly is not observable at the surface, is an example of an internal wave.

The restoring force for waves is proportional to the product of gravity and the density difference between the two layers (the relative buoyancy). At internal interfaces this difference is much smaller than the density difference between air and water (by several orders of magnitude). As a consequence, internal waves can attain much larger amplitudes than surface waves. It also takes longer for the restoring force to return particles to their average position, and internal waves have periods much longer than surface gravity waves (from 10 - 20 minutes to several hours, compared to several seconds or minutes for surface gravity waves). In contrast to surface waves in which horizontal particle velocities are largest at the surface and either decay quickly with depth (in deep water waves) or are independent of depth (in shallow water waves), horizontal water movement in internal waves is largest near the surface and bottom and minimal at mid-depth.

Internal waves can often be observed in the atmosphere, where they travel on the interface between warm and cold air. Figure 10.5 and Figure 10.6 show two examples.

Figure 10.5. Cloud patterns in which clouds are organised in bands are not an unusual sight. The photo shows a cloud pattern produced by an internal wave. The air rises with the crest of the wave, cooling in the process, which produces condensation (cloud formation). It sinks in the troughs, causing warming and evaporation of the cloud moisture. The wave crests then become visible as cloud bands.
The photo was taken near Mount Lawson (30 km west of Brisbane, Australia) and the wave was travelling towards Brisbane city, while it was a fine still day on the ground.

Figure 10.6. A particularly spectacular cloud pattern produced by a breaking internal wave in the atmosphere over the Black Forest, Germany.

Figure 10.7 shows an example of an internal wave travelling on the seasonal thermocline in coastal waters. Such waves typically have wave lengths of several tens of metres and periods of about 30 minutes. Convergence of surface particle movement above the wave troughs near the surface often collects floating matter and makes the waves visible as slick marks (Figure 10.8). If the interface on which the wave travels is very shallow, ships may find themselves in a situation that most of the energy put into the propeller goes into driving the circular particle motion of the internal wave at the interface, with the ship making little or no progress through the water. This phenomenon is known as "dead water" and is common in fjords, where the interface is produced by a shallow layer of freshwater from glacier runoff overlying oceanic water underneath.

The most common internal waves are of tidal period and manifest themselves in a periodic lifting and sinking of the seasonal and permanent thermocline at tidal rythm. In some ocean regions their surface expressions, produced by convergence over the wave troughs, is visible in satellite images (Figure 10.9).

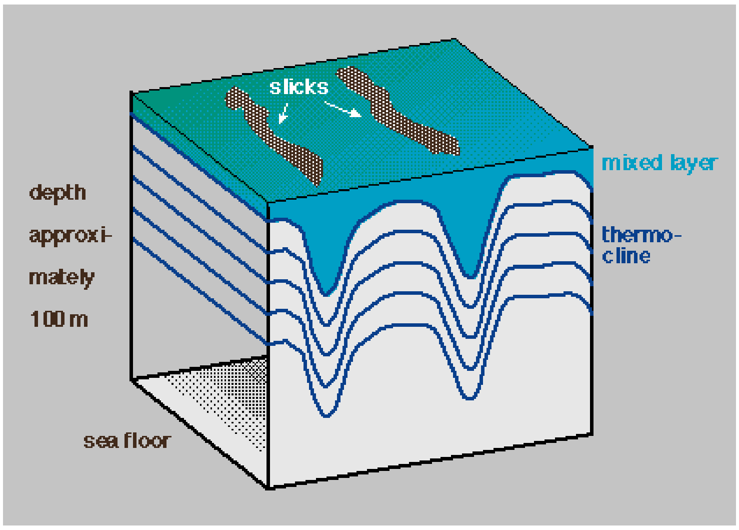


Figure 10.7. Sketch of an internal wave propagating on the seasonal thermocline in the coastal ocean. Blue contours indicate isotherms (contours of constant temperature). The mixed layer is indicated by the greenish-blue region above the first isotherm, the thermocline by the crowding of isotherms at mid-depth. The slicks on the surface are

produced by the convergence of water above the wave troughs in the mixed layer. Total water depth of the example shown is approximately 100 m.

Figure 10.8. An internal wave travelling up the Derwent River estuary in Hobart, Tasmania. The effect of the wave is visible by streaks of smooth water produced by the convergences above the wave troughs (compare Figure 10.7).The wave seen here is of the shortest possible wavelength (compare the distance between the streaks with the yacht) and has a period of 10 - 20 minutes. The estuary stretches from left to right. Note how the shape of the streaks reflects the faster propagation speed of the internal wave over deeper water (in the central channel of the estuary). The photograph was taken at the CSIRO Marine Laboratories.

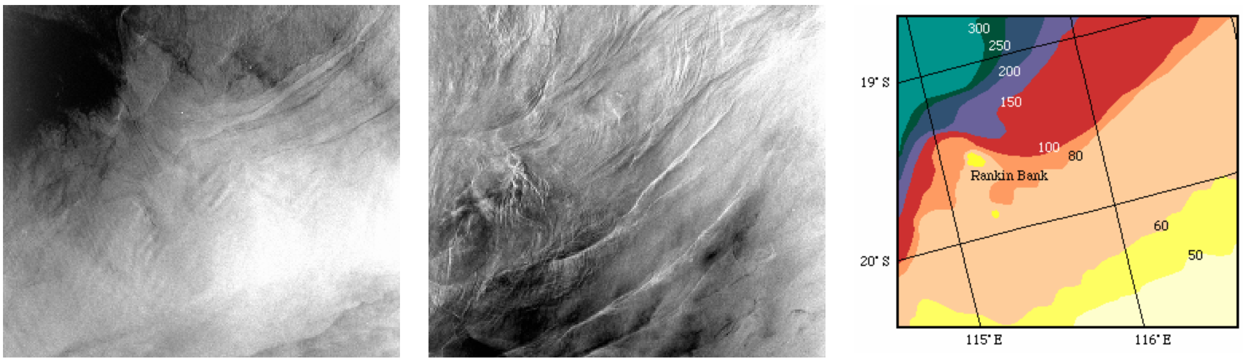


Figure 10.9. Surface expressions of internal tides in Canadian Space Agency/Agence Spatiale Canadienne RADARSAT Synthetic Aperture Radar (SAR) scenes from the Australian North-West Shelf.
The North-West Shelf experiences some of the largest tides in the world ocean, with a tidal range of over 8 m. An internal tide is produced as the surface tide impinges on the continental shelf. It travels on the thermocline from the shelf break towards the coast.
The images are approximately 200 km x 200 km in size; north is up. The coast is towards the lower right outside the image range. The figure on the right shows the orientation of the image and the topography of the region; depth contours are given in metres.
The waves are observable in the RADARSAT images as bands of alternating rough and smooth surface, the effect of current convergence on the wind waves and accumulation of floating planctonic matter. The images show bands of semi-circular shape, produced by different propagation speed of the internal waves. Some bands are nearly straight indicating uniform propagation speed in these regions. The effect of the shallow Rankin Bank on the propagation of the waves is clearly visible. © 1998 Canadian Space Agency/Agence Spatiale Canadienne RADARSAT ADRO Project #72; Cresswell and Tildesley, reproduced by permission.

Lecture 11

Tides

reviewed by: John Luick

Tides are long waves, either progressing or standing. The dominant period usually is 12 hours 25 minutes, which is 1/2 of a lunar day. Tides are generated by the gravitational potential of the moon and the sun. Their propagation and amplitude are influenced by friction, the rotation of the earth (Coriolis force), and resonances determined by the shapes and depths of the ocean basins and marginal seas.

The most obvious expression of tides is the rise and fall in sea level. Equally important is a regular change in current speed and direction; tidal currents are among the strongest in the world ocean.

Description of tides

- High water: a water level maximum ("high tide")
- Low water: a water level minimum ("low tide")
- Mean Tide Level: the mean water level, relative to a reference point (the "datum") when averaged over a long time
- Tidal range: the difference between high and low tide
- Daily inequality: the difference between two successive low or high tides
- Spring tide: the tide following full and new moon
- Neap tide: the tide following the first and last quarter of the moon phases.

The result of alternate spring and neap tides is a half monthly inequality in tidal heights and currents. Its period is 14.77 days, which is half the synodic month. (Synodic: related to the same phases of a planet or its satellites. A synodic period or synodic month is thus the time that elapses between two successive identical phases. In tidal theory synodic always refers to the moon, so a synodic month is the time that elapses between successive identical phases of the moon, for example between successive new moons.) There are other inequalities with similar and longer periods.

The Tide-Generating Forces

As the earth revolves around the gravitational centre of the sun/earth system, the orientation of the earth´s axis in space remains the same. This is called **revolution without rotation**.

The tide generating force is the sum of gravitational and centrifugal forces. In revolution without rotation the centrifugal force is the same for every point on the earth´s surface, but the gravitational force varies (Figure 11.1).

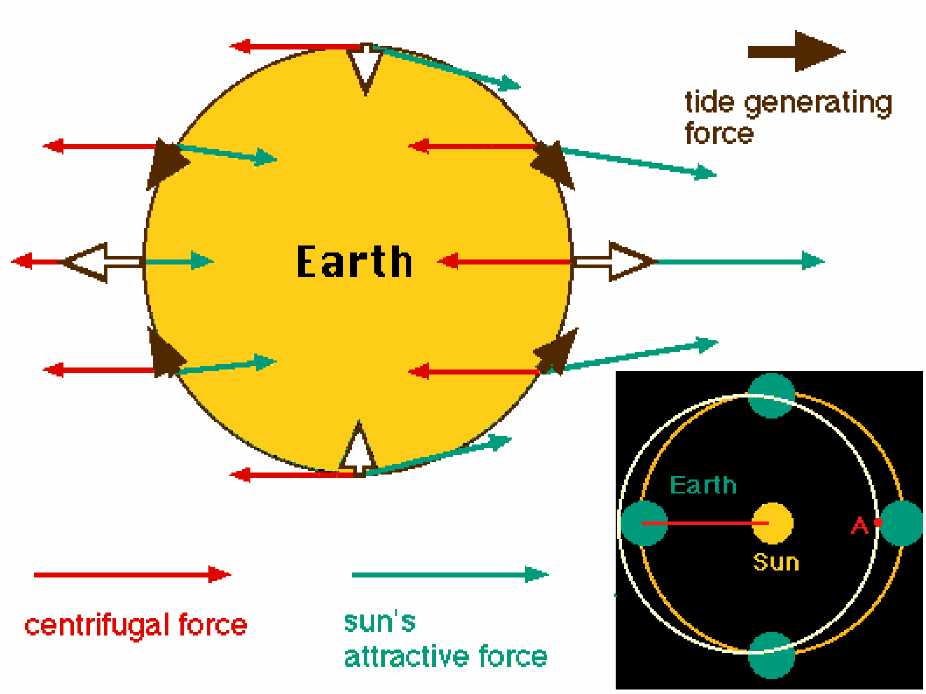


Figure 11.1. The tide generating force as the resultant of centrifugal and gravitational forces.
The figure on the right shows the earth's movement as revolution without rotation. The golden circle shows the path of the earth's centre in space, the white circle the path of point **A**. Note that the orientation of the earth's axis in space does not change and that as a consequence the diameter of both circles is the same. This means that the centrifugal force experienced by all points on the earth (as well as inside it) is the same, in magnitude and direction.

The gravitational force exerted by the sun always points to the centre of the sun. The effect of this force experienced by points on the earth's surface therefore varies with position, in magnitude and direction. The resulting balance of forces is shown in the left diagram. The open arrows indicate a net force in the vertical direction, the full arrows indicate a net force that contains also a horizontal component. This horizontal component of the resulting force is the tide generating force.

The same principle applies to the interaction between the earth and its moon. Both bodies revolve around their common centre of gravity, which in this case is inside the earth (but not at its centre). The earth again revolves around this centre without rotation, so that the centrifugal force is the same everywhere but the gravitational force exerted by the moon varies over the earth's surface.

Note again that in addition to revolving around the sun without rotation, the earth spins around its axis. This rotation around its own axis is an entirely different issue and does not invalidate the findings about the balance between gravitational and centrifugal force with respect to the earth's revolution around the sun. Its only effect on the tides is that it moves the entire tide-generating force field (shown above) around the earth once every day.

It follows that the tide generating force varies in intensity and direction over the earth's surface. Its vertical component is negligibly small against gravity; its effect on the ocean can be disregarded. Its horizontal component produces the tidal currents, which result in sea level variations (Figure 11.2).

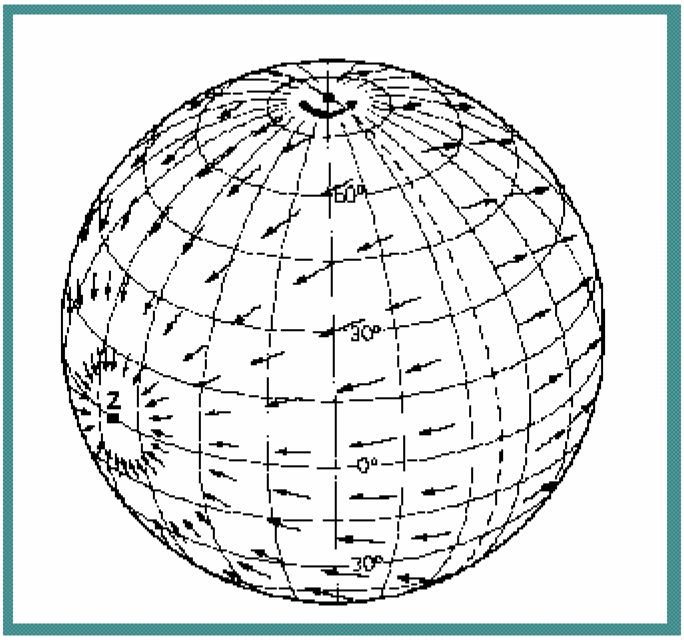


Figure 11.2. A snapshot of the tide generating force when the moon is over the point marked Z (the "zenith"). This force field rotates around the earth with the moon. Note that there are two points of water accumulation (high water), ie the basic tidal period is half the period of the moon's revolution.

The gravitational force exerted by a celestial body (moon, sun or star) is proportional to its mass but inversely proportional to the square of the distance. The Sun's mass is equivalent to some 332,000 Earth masses, while the mass of the Moon corresponds to only 1.2 percent of the mass of the Earth. The mean distance Sun -Earth is 149.5 million km, the mean distance Earth - Moon only 384,000 km. If the gravitational force of the Sun and Moon are compared, it is found that the Sun's enormous mass easily makes up for its larger distance to Earth, to the extent that the gravitational force of the Sun felt on Earth is about 178 times that of the Moon. As a result the Earth's orbit around the Sun is not seriously distorted by the Moon's movement around the Earth.

However, as is evident from Figure 11.1, tides are not produced by the absolute pull of gravity exerted by the Sun and the Moon but by the differences in the gravitional fields produced by the two bodies across the Earth's surface. Because the Moon is so much closer to the Earth than the Sun, its gravitational force field varies much more strongly over the surface of the Earth than the gravitational force field of the Sun. Quantitative analysis shows that the *differences* of the gravitational forces across the Earth's surface are proportional to the *cube* of the distances Sun - Earth and Earth - Moon. As a result the Sun's tide-generating force is only about 46% of that from the Moon. Other celestial bodies do not exert a significant tidal force.

Main tidal periods

- Tides produced by the moon
  - M_2 (semidiurnal lunar) 1/2 lunar day = 12h 25min
  - O_1 (diurnal lunar) 1 lunar day = 24h 50min
- Tides produced by the sun
  - S_2 (semidiurnal solar) 1/2 solar day = 12h
  - K_1 (diurnal solar) 1 solar day = 24h

The tides can be represented as the sum of harmonic oscillations with these periods, plus harmonic oscillations of all the other combination periods (such as inequalities). Each oscillation, known as a tidal constituent, has its amplitude, period and phase, which can be extracted from observations by harmonic analysis. Hundreds of such oscillations have been identified, but in most situations and for predictions over a year or so it is sufficient to include only M_2 , S_2 , K_1 and O_1. Practical predictions produced on computers for official tide tables use significantly more terms than these four; for example, the Australian National Tidal Facility uses 115 terms to produce the official Australian Tide Tables.

Tidal Classification

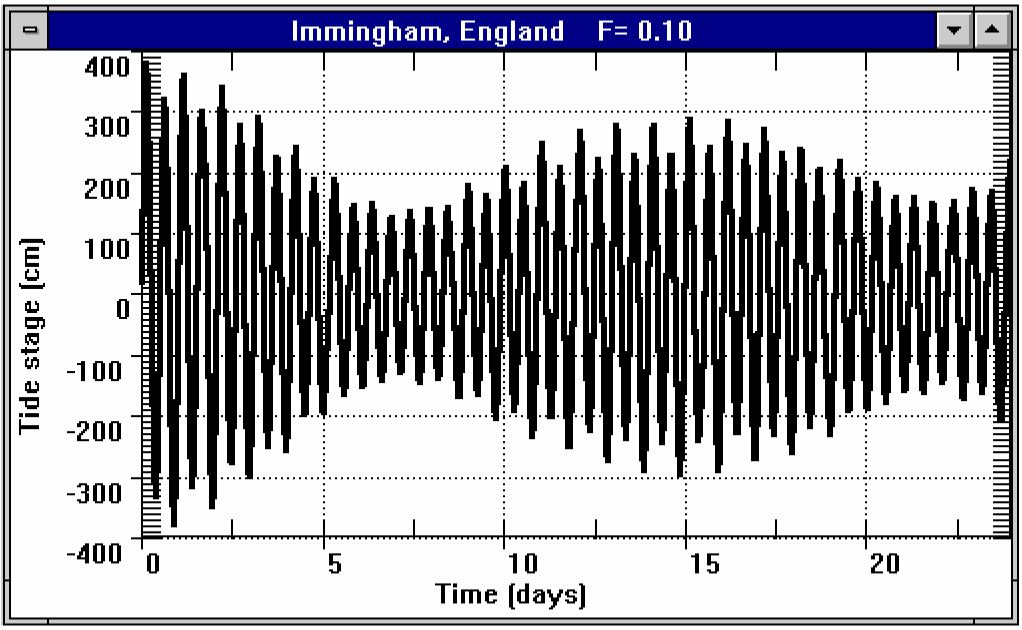
The form factor F is used to classify tides. It is defined as

$F = (K_1 + O_1) / (M_2 + S_2)$

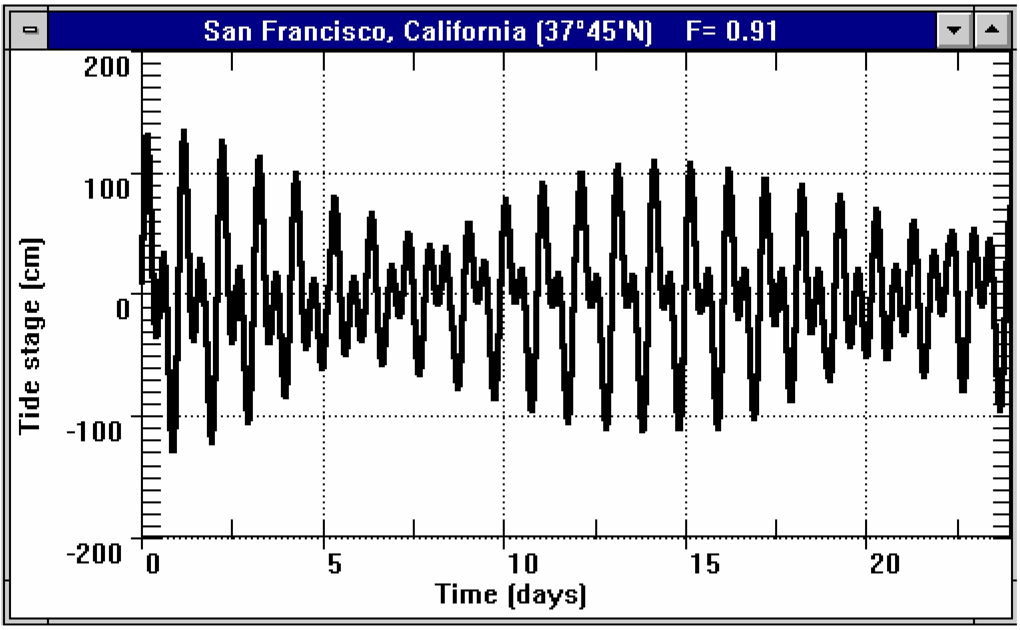
where the symbols of the constituents indicate their respective amplitudes. Four categories are distinguished :

value of F	category
0 - 0.25	semidiurnal
0.25 - 1.5	mixed, mainly semidiurnal
1.5 - 3	mixed, mainly diurnal
> 3	diurnal

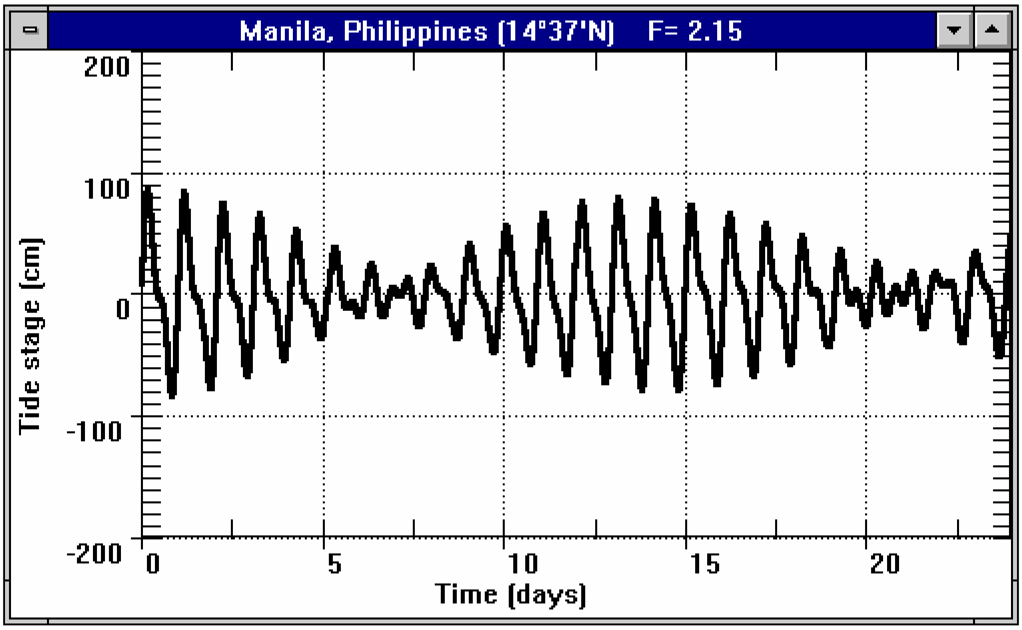
Figure 11.3 shows examples.



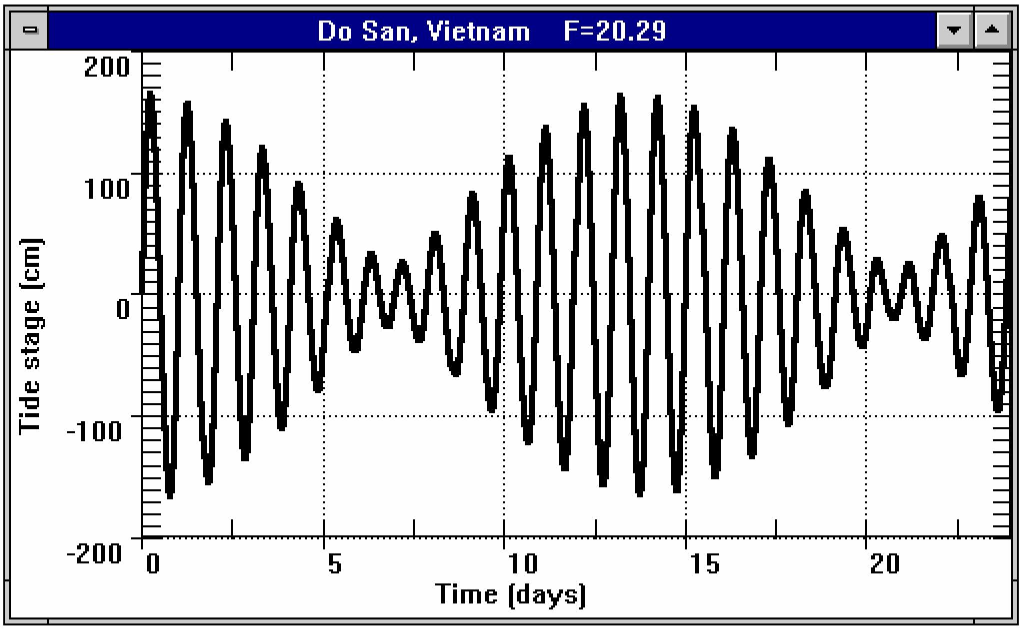
Immingham: semidiurnal; two high and low waters each day.



San Francisco: mixed, mainly semidiurnal; two high and low waters each day during most of the time, only one high and low water during neap tides.



Manila: mixed, mainly diurnal, one dominant high and low water each day, two high and low waters during spring tide.



Do San: diurnal; one high and low water each day.
Figure 11.3. Sea level as function of time at four ports (above).

Shape of the Tidal Wave

The scales of variations in the forcing field are of global dimensions. Only the largest water bodies can accommodate directly forced tides. On a non-rotating earth the tides would be standing waves; they would have the form of seiches, ie a back and forth movement of water across lines of no vertical movement (nodes). On a rotating earth the tidal wave is transformed into movement around points of no vertical movement known as **amphidromic points**.

- At amphidromic points the tidal range is zero.
- Co-range lines (lines of constant tidal range) run around amphidromic points in quasi-circular fashion.
- Co-phase lines (lines of constant phase, or lines which connect all places where high water occurs at the same time) emanate from amphidromic points like spokes of a wheel.

The animation compares seiche movement with tidal movement around an amphidromic point. Note that on a rotating earth the tides take the shape of propagating waves: The wave propagates around the amphidromic point in clockwise or anti-clockwise fashion.

Details of the shape of the tidal wave depend on the configuration of ocean basins and are difficult to evaluate. Computer models can give a description of the wave on an oceanic scale (Figure 11.4). Their results have to be verified against observations of tidal range and times of occurrence of high and low water. Distortions of the wave on the continental shelf caused by shallow water make it difficult to assess results for the open ocean. In deep water the tidal range rarely exceeds 0.5 m.

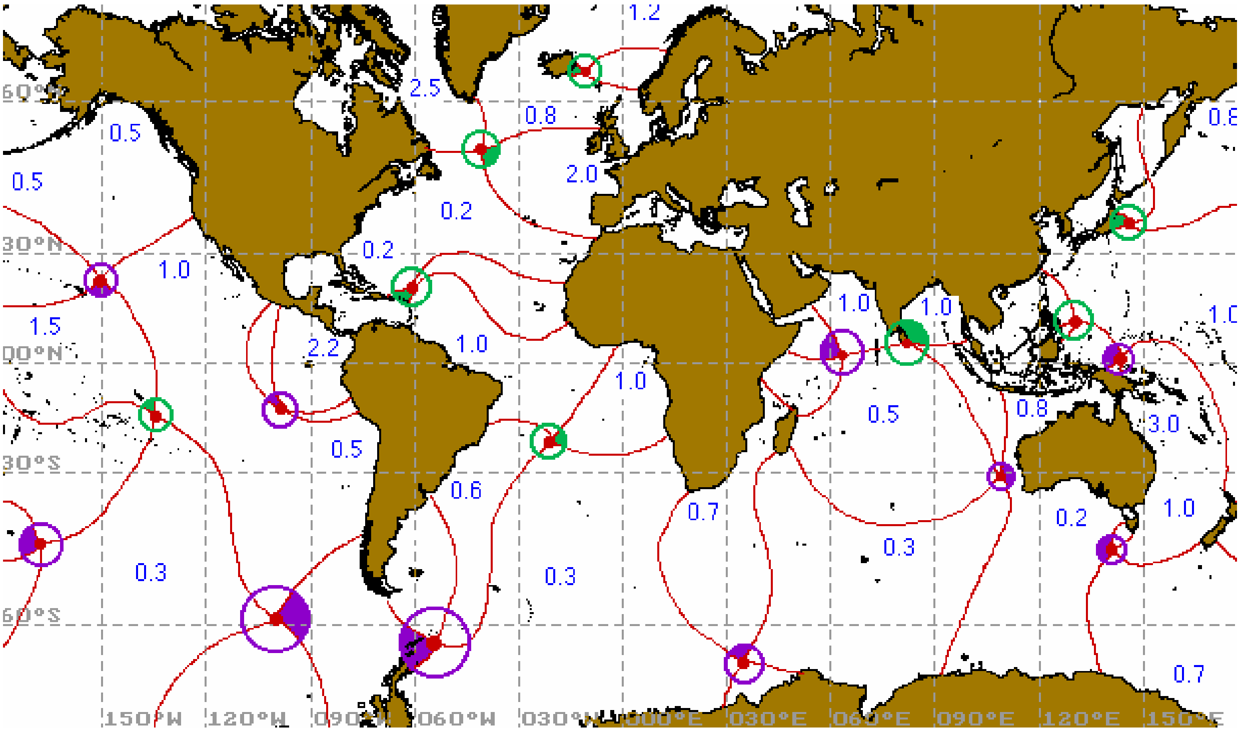


Figure 11.4. An example of a result from a computer model of the tides in the world ocean, showing amphidromic points (red circles) and sense of rotation (purple = clockwise, green = anti-clockwise). The figure shows the M_2 wave. The complete solution consist of superposition of this result with the results for S_2, K_1 and O_1 (and other constituents).
Red lines are co-phase lines indicating the crest of the tidal wave at 0, 3, 6 and 9 lunar hours after the moon's transit through the meridian of Greenwich. (The filled sector of the purple and green circles indicates the sector between 0 and 3 lunar hours; 12 lunar hours correspond to 12 hours 25 minutes.)
Blue numbers indicate the tidal range (m) due to the M_2 tide. In the open ocean the tidal range is generally small (0.2 - 0.5 m; it is zero at amphidromic points). The model indicates large tidal ranges closer to the coast; but the coarse model resolution does not reproduce the known extreme tides in the Bay of Fundy or on the French coast. (These regions show up with very large tidal ranges, though.)

Co-oscillation tides

Tides in marginal seas and bays cannot be directly forced; they are co-oscillation tides generated by tidal movement at the connection with the ocean basins. Depending on the size of the sea or bay they take the shape of a seiche or rotate around one or more amphidromic points.

If the tidal forcing is in resonance with a seiche period for the sea or bay, the tidal range is amplified and can be enormous. This produces the largest tidal ranges in the world ocean (14 m in the Bay of Fundy on the Canadian east coast; 10 m at St. Malo in France, 8 m on the North West Shelf of Australia and at the extreme north of the Gulf of California in Mexico; all are mainly semidiurnal tides). Tidal range is then largest at the inner end of the bay, in accordance with the dynamics of seiches in open basins. Modest amplification is exerienced in Spencer Gulf of South Australia where the tidal range at spring tide is 3 m at the inner end, while it is less than 1 m at the entrance to the Gulf.

Figure 11.5 shows an example of a co-oscillation tide in a large bay. The tide is forced from the open end by the oceanic tide, which has an maximum tidal range (at spring tide) of about 1 m. Because of the width of the basin the Coriolis force is able to shape the wave, producing amphidromic points around which the wave propagates. Amplification is particularly large along the British coast and in the English Channel.

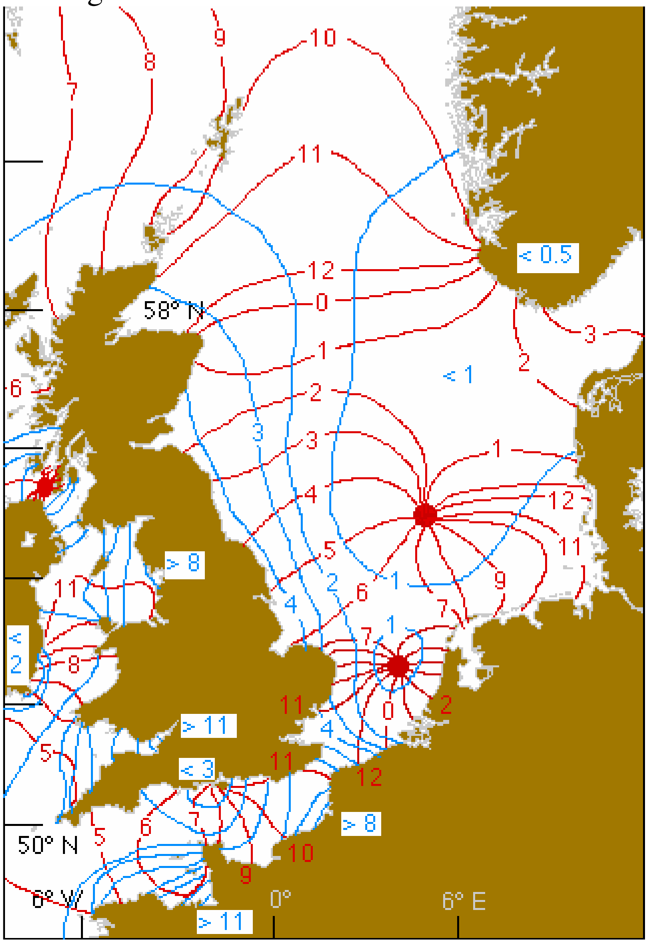


Figure 11.5. Tides in the North Sea as derived from observations. Red lines are co-phase lines of the M_2 tide, labelled in hours after the moon's transit through the meridian of Greenwich. (There are thus only 25 minutes between the co-phase lines labelled 12 and 0.) Blue lines give the mean tidal range at spring tide (co-range lines of the sum of M_2 and S_2).

The progress of the tidal wave from the Atlantic Ocean into the North Sea is clearly demonstrated by the co-phase lines. The wave enters from the north and propagates along the British coast; it then proceeds around two amphidromic points along the Dutch, German and Danish coastline. Another wave enters from the south west, through the English Channel. In the Irish Sea the wave enters from the south.

The influence of the Coriolis force is demonstrated by the co-range lines, which show large tidal range along the British coast and small tidal range along the German, Danish and Norwegian coast. The same effect (amplification on the right side of the wave) is seen in the English Channel, where the tidal range along the French coast is as high as 11 m compared with 3 m on the English coast, and in the Irish Sea, where 8 m on the English coast compare with 2 m on the Irish coast.

Lecture 12

Estuaries

reviewed by: Gunther Krause

Estuaries are regions of the coastal ocean where salinity variations in space are so large that they determine the mean circulation. They are related to mediterranean seas in the sense that the major driving forces for their circulation are thermohaline processes. They differ from mediterranean seas mainly in size and configuration. Most estuaries are found at river mouths; they are thus long and narrow, resembling a channel. Compared to the flow in the direction of the estuary axis, cross-channel motion is very restricted, and the estuarine circulation is well described by a two-dimensional current structure. This is not true for mediterranean seas, which are wide enough to accommodate flow in three directions and allow the Coriolis force to exert its influence (e.g. in inertial motion). Modification of the circulation by winds is also stronger in mediterranean seas than in estuaries, where the circulation is restricted to the direction of the estuary axis regardless of winds.

A classical definition of estuaries used by the United Nations Educational, Scientific and Cultural Organization (UNESCO) and reproduced in textbooks is

a semi-enclosed coastal body of water having free connection to the open sea and within which sea water is measurably diluted with fresh water deriving from land drainage.

This definition works well for estuaries in the temperate zone where estuaries are linked with river mouths but does not include bodies of anomalously high salinity such as lagoons, or coastal inlets which are connected to the ocean only occasionally. For Australian (and indeed world-wide) application it is advisable to amend the definition to:

An estuary is a semi-enclosed coastal body of water having free connection to the open sea at least intermittently, and within which the salinity is measurably different from the salinity in the adjacent open sea.

Estuaries can be grouped into classes, according to their circulation properties and the associated steady state salinity distribution. The most important estuary types are

1. salt wedge estuary
2. highly stratified estuary
3. slightly stratified estuary
4. vertically mixed estuary
5. inverse estuary
6. intermittent estuary

The following discussion concentrates on types 1-4 and closes with some remarks on 5-6.

The balance of forces that establishes a steady state in types 1-4 involves advection of freshwater from a river and introduction of sea water through turbulent mixing. Mixing is achieved by tidal currents. (This is another aspect where estuaries differ from mediterranean seas; mixing in mediterranean seas is usually associated with eddies but not with tidal currents, which in most mediterranean seas are quite small.) The ratio of freshwater input to sea water mixed in by the tides determines the estuary type. One way of quantifying this is by comparing the volume R of freshwater that enters from the river during one tidal period, with the volume V of water brought into the estuary by the tide and removed over each tidal cycle. R is sometimes called the river volume, while V is known as the tidal volume. It is important to note that it is only the ration R : V that determines the estuary type, not the absolute values of R or V. In other words, estuaries can be of widely different size and still belong to the same type. Salt wedge estuaries for example can be produced by a small creek in a nearly tide-free bay, or they can be of the scale of the Mississippi and Amazon rivers, which carry so much water that even strong tidal mixing is insignificant in comparison.

Salt wedge estuary

River volume R is very much larger than the tidal volume V, or there are no tides at all. The fresh water flows out over the sea water in a thin layer. All mixing is restricted to the thin transition layer between the fresh water at the top and the "wedge" of salt water underneath. Vertical salinity profiles therefore show zero salinity at the surface and oceanic salinity near the bottom all along the estuary. The depth of the interface decreases slowly as the outer end of the estuary is approached (Figure 12.1). Examples of large salt wedge estuaries are the Mississippi and the Congo Rivers. Other examples may be as small as only a few kilometers long. Note the vertical exaggeration in this and the following figures.

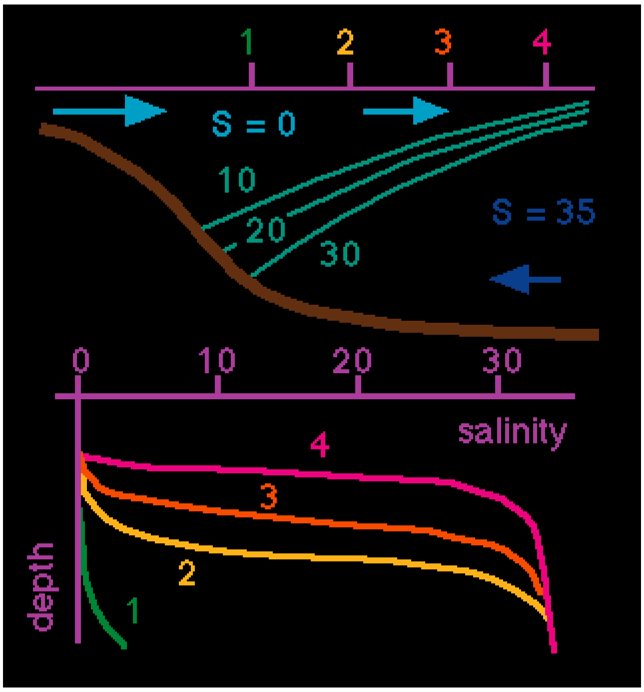


Figure 12.1. Salinity in a salt wedge estuary. Top: as a function of depth and distance along the estuary, numbers indicate station locations; bottom: in vertical salinity profiles for stations 1 - 4. Surface salinity is close to zero at all stations, bottom salinity close to oceanic.

Highly stratified estuary

River volume R is comparable to but still larger than tidal volume V. Strong velocity shear at the interface produces internal wave motion at the transition between the two layers. The waves break and "topple over" in the upper layer, causing entrainment of salt water upward. Entrainment is a one-way process, so no fresh water is mixed downward. This results in a salinity increase for the upper layer, while the salinity in the lower layer remains unchanged, provided the lower layer volume is significantly larger than the river volume R and can sustain an unlimited supply of salt water (Figure 12.2). Examples of this type of estuary are fjords, which are usually very deep and have a large salt water reservoir below the upper layer. Moderately deep river beds often exhibit this type of stratification during periods of weak river flow.

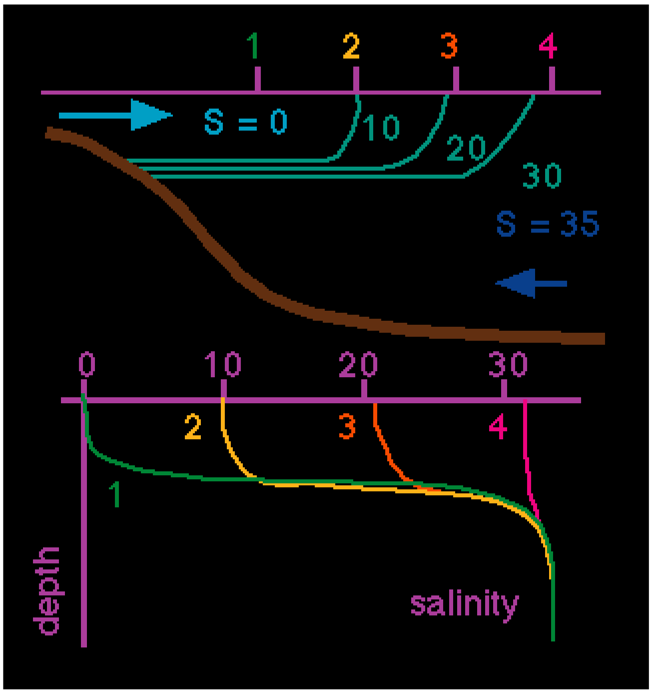


Figure 12.2. Salinity in a highly stratified estuary. Top: as a function of depth and distance along the estuary, numbers indicate station locations; bottom: in vertical salinity profiles for stations 1 – 4. Surface salinity increases from station 1 to station 4, but bottom salinity is close to oceanic at all stations.

The upward mass flux of salt water leads to an increase of flow speed in the upper layer. This increase of mass transport in the upper layer can be quite significant, to the extent that the river output appears insignificant compared with the overall circulation (Figure 12.3). A 20-fold amplification of the mass transport into the sea is quite realistic. The surface velocity increases likewise, although not as dramatically, as the downstream increase in width of the estuary compensates for some of the increase in mass transport.

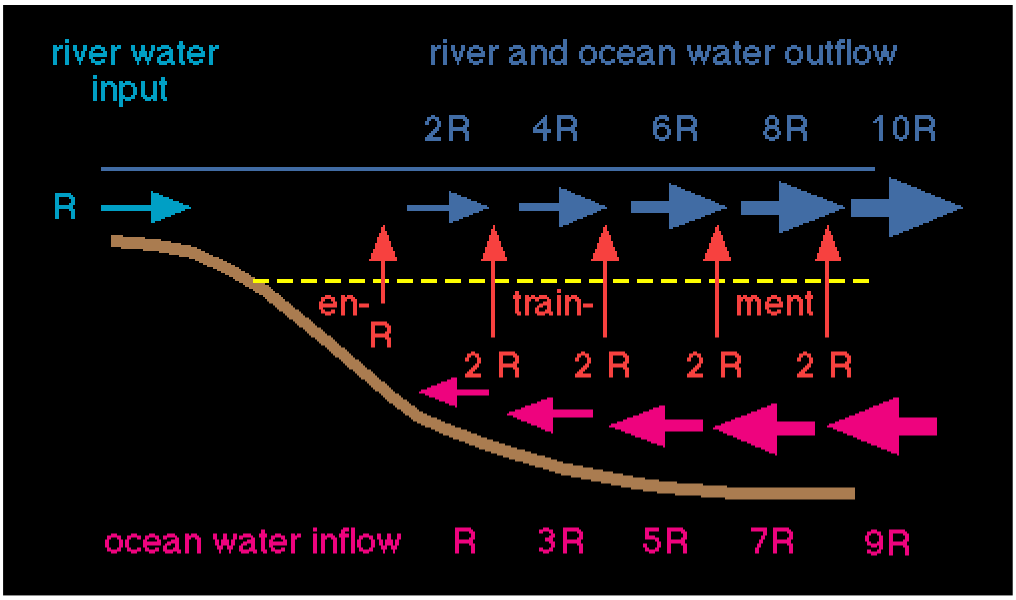


Figure 12.3. Sketch of mass transport in a highly stratified estuary. River volume input is R. Outflow from the estuary in the upper layer is 10R; this is compensated by inflow of oceanic water of 9R. The net outflow at the outer end of the estuary is of course still only 1R.

Slightly stratified estuary

River volume *R* is small compared to tidal volume *V*. The tidal flow is turbulent through the entire water column (the turbulence induced mainly at the bottom). As a result, salt water is stirred into the upper layer and fresh water into the lower layer. Salinity therefore changes along the estuary axis not only in the upper layer (as was the case in the highly stratified estuary) but in both layers (Figure 12.4). There is some increase in surface velocity and upper layer transport towards the sea but not nearly as dramatic as in the highly stratified case. This type of estuary is widespread in temperate and subtropical climates; many examples are found around the world.

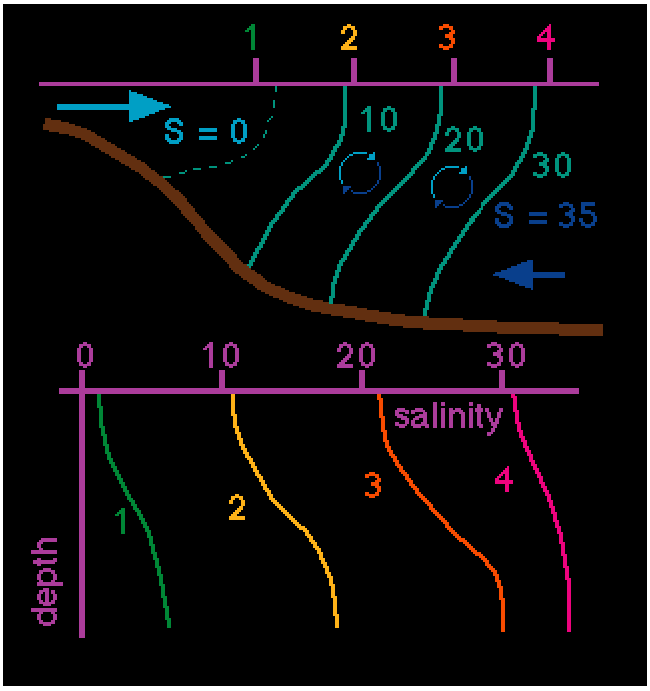


Figure 12.4. Salinity in a slightly stratified estuary. Top: as a function of depth and distance along the estuary, numbers indicate station locations, mixing between upper and lower layer is indicated by the faint arrows; bottom: in vertical salinity profiles for stations 1 - 4. Surface and bottom salinity increase from station 1 to station 4, but surface salinity is always lower than bottom salinity.

Vertically mixed estuary

River volume *R* is insignificant compared with tidal volume *V*. Tidal mixing dominates the entire estuary. Locally it achieves complete mixing of the water column between surface and bottom, erasing all vertical stratification. As a result, vertical salinity profiles show uniform salinity but a salinity increase from station to station as the outer end of the estuary is approached (Figure 12.5). This type of estuary is found in regions of particularly strong tides; an example is the River Severn in England.

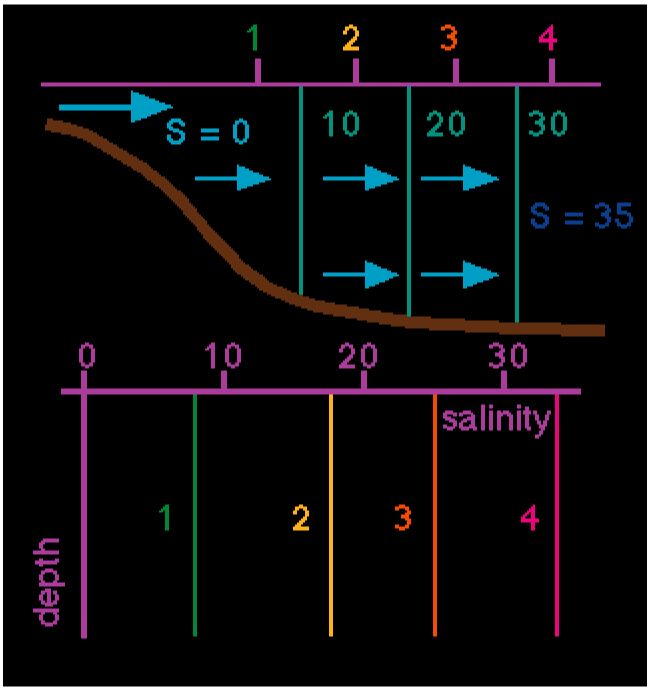


Figure 12.5. Salinity in a vertically mixed estuary. Top: as a function of depth and distance along the estuary, numbers indicate station locations; bottom: in vertical salinity profiles for stations 1 - 4. Surface and bottom salinity increase from station 1 to station 4, but surface salinity is always nearly identical to bottom salinity.

As said above, the estuary type is determined by the ratio *R* : *V*. Varying this ratio produces a range of salinity distributions, which can be classified by the ratio of surface salinity S_s against bottom salinity S_b. The ratio S_s : S_b can therefore be used as a substitute for the ratio *R* : *V*. Salinities are easier to measure than tidal and river volume, and a ratio based on salinities is therefore of practical value. Figure 12.6 shows the unified classification scheme based on salinity. The salt wedge estuary has freshwater at the surface, oceanic water at the bottom and is thus identified by a salinity ratio of zero. It occupies the bottom line of the diagram. Salinity in the vertically mixed estuary varies along the estuary but is the same at the surface and at the bottom everywhere, so the vertically mixed estuary has a salinity ratio of one and occupies the top line of the diagram. The highly stratified estuary is found in the lower right triangle, the slightly stratified estuary in the upper left triangle.

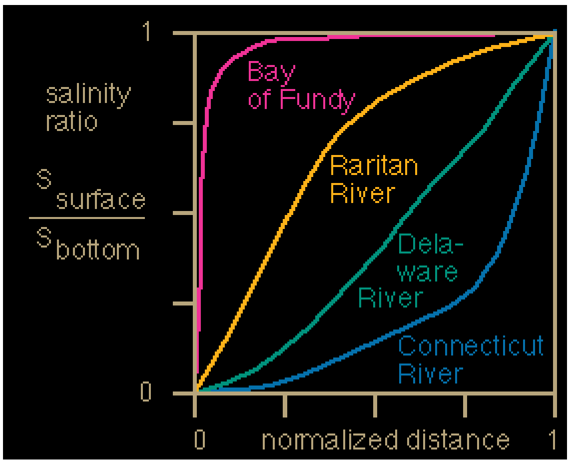


Figure 12.6. Classification diagram for estuaries based on the ratio surface salinity : bottom salinity, with examples for different estuary types. Normalized distance is the distance along the estuary, divided by its length. The

Connecticut River is a salt wedge estuary, The Delaware River highly stratified, the Raritan River slightly stratified, and the Bay of Fundy vertically mixed.

Estuaries can change type as a result of variations in rainfall and associated river flow. They may show different characteristics in different parts as a result of topographic restrictions on the propagation of the tide along the estuary which affect the tidal volume. The classification diagram can be used to establish changes of estuary type in space and time.

Inverse estuaries

These estuaries have no fresh water input from rivers and are in a region of high evaporation. Surface salinity does not decrease from the ocean to the inner estuary, but water loss from evaporation leads to a salinity increase as the inner end of the estuary is approached (Figure 12.7). This results in a density increase and sinking of high salinity water at the inner end. As a result, movement of water is directed inward at the surface and towards the sea at the bottom, with sinking at the inward end. Compared to the estuaries discussed above their circulation is reversed, which explains the name inverse estuary.

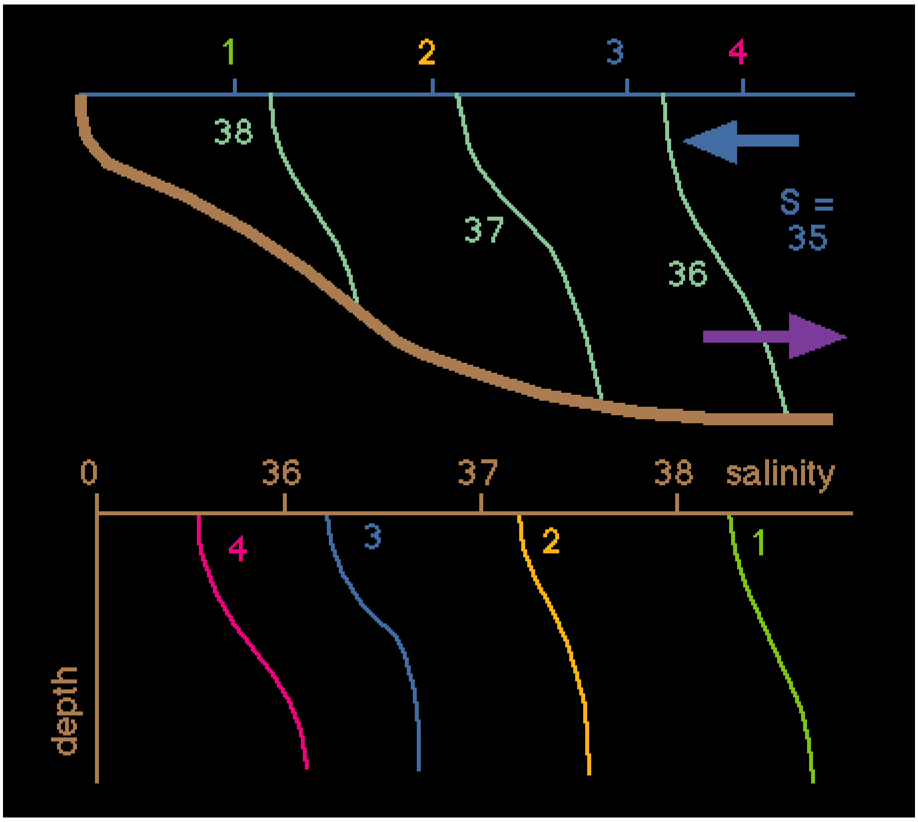


Figure 12.7. Salinity in an inverse estuary. Top: as a function of depth and distance along the estuary, numbers indicate station locations; bottom: in vertical salinity profiles for stations 1 - 4. The circulation is into the estuary at the surface; outflow occurs at depth. Surface and bottom salinities decrease from station 1 to station 4, but surface salinity is always lower than bottom salinity.

Some tropical Australian estuaries show a combination of 'normal' and 'inverse' circulation. Figure 12.8 shows the example of the Alligator River. The estuary receives some fresh water from river inflow but evaporation is so strong that at some intermediate position all river water has evaporated and salinity becomes higher than in the open sea. Upstream from this position the circulation is "normal"; downstream it is "inverse". Other examples are the Escape River and the Wenlock and Duncie Rivers.

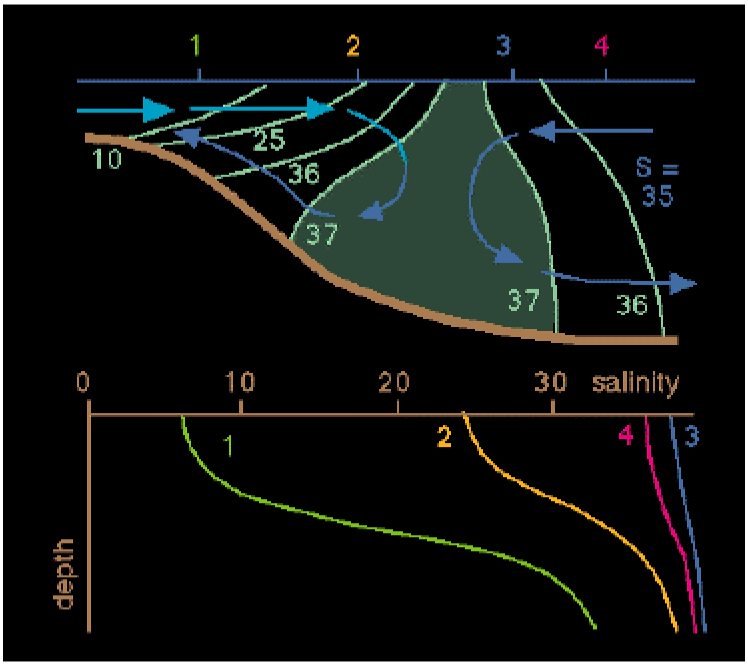


Figure 12.8. Circulation in the Alligator River (Northern Territory), showing a combination of a slightly stratified estuary in the inner estuary (stations 1 and 2) with an inverse estuary in the outer region (stations 3 and 4). Note the salinity maximum known as the "salinity plug", between stations 2 and 3. It is produced by the evaporation over the region.

Intermittent estuaries

Many estuaries change their classification type because of highly variable rainfall over the catchment area of their river input. River input may be small, but as long as some fresh water enters the estuary, the estuarine character is maintained (in the form of a salt wedge estuary). If the river input dries up completely during the dry season, estuaries loose their identity and turn into oceanic embayments. An example is the South West Arm of Port Hacking south of Sydney which turns into a highly stratified estuary for a few weeks after heavy rains.

During their estuarine periods intermittent estuaries can be classified on the basis of the classification diagram of Figure 12.6, but the effect of their high environmental variability on marine life is so overwhelming that a separate classification appears justified. Marine life in intermittent estuaries undergoes a complete change of communities between the estuarine and oceanic phases - very few plants or animals can cope with the salinity changes that occur between the two phases.

Lecture 13

Oceanographic instrumentation

reviewed by: Ian Helmond

Physical oceanography is an experimental science and requires observation and exact measurement to achieve its aims. It can draw on the experience of the related science fields of physics and chemistry and make use of existing achievements in the areas of technology and engineering, but the oceanic environment places unique demands on instrumentation that are not easily met by standard laboratory equipment. As a consequence, the development and manufacturing of oceanographic instrumentation developed into a specialised activity. Manufacturers of oceanographic equipment serve a small market but aim at selling their products world-wide.

This lecture gives an overview of the range of instrumentation used at sea and the principles involved. It aims at covering standard classical instruments as well as modern developments. The following table summarises its content.

research need	**available equipment / instrumentation**
provision of observing platform	• research vessels • moorings • satellites • submersibles • towed vehicles • floats and drifters
measurement of hydrographic properties (temperature, salinity, oxygen, nutrients, tracers)	• reversing thermometers • Nansen and Niskin bottles • CTDs • multiple water sample devices • thermosalinographs • remote sensors
measurement of dynamic properties (currents, waves, sea level, mixing processes)	• current meters • wave measurements • tide gauges • remote sensors • shear probes

Platforms

All measurements at sea require a reasonably stable platform to carry the necessary instrumentation. The platform can be at the sea surface, at the sea floor, in the ocean interior or in space. The choice of platform depends on its capabilities to collect the required information in space and time.

Research vessels

Like any other ocean going ship, research vessels have to meet the requirement to be seaworthy and capable of riding out bad weather. The weather conditions in the investigation area thus define the minimum size for the vessel. Additional requirements, such as the handling of heavy equipment at sea or the need for a large scientific party during an interdisciplinary study, can increase the minimum size. Typical ocean going research vessels are 50 - 80 m long, have a total displacement of 1000 - 2000 tonnes and provide accommodation for 10 - 20 scientists (Figure 13.1).

Figure 13.1. The Australian research vessel *Franklin* in Darwin harbour. The accommodation and laboratories are in the front and centre, with the working deck area towards the stern. A large A-frame is located at the stern, a smaller A-frame for CTD stations midships on the side. The bridge house wraps around one side of the vessel, allowing the officers on the bridge to watch the activity on the working deck and at the CTD when the ship is on station.

The shape of a research vessel is determined by the need for a reasonably large working deck, several powerful winches for lowering and retrieving instrumentation and at least one "A-frame", a structure which allows a wire to go from the ship's winch over the side of the ship or over the stern and vertically into the water (Figure 13.2).

Figure 13.2. Part of the working deck with the A-frame at the stern of the Australian research vessel *Franklin*, seen during the deployment of a surface buoy for meteorological and oceanographic measurements.

The requirement to be at sea for extensive periods of time, remain stationary while equipment is handled over the side and move at very slow speed when equipment is towed behind the vessel place additional demands on research vessel design. To increase endurance (the number of days a ship can remain at sea before running out ouf fuel), research vessels have only a moderate operating speed of 10 - 12 knots (18 - 28 km/hr). This compares with operating speeds of 15 - 20 knots for merchant ships. Most research vessels have an endurance of 20 - 25 days, which gives them a range of 6000 - 8000 nautical miles (11,000 - 14,800 km), sufficient to operate at the high seas within a few days of reach of land. Only the major oceanographic institutions of the world operate research vessels with global research capability.

All modern ships are powered by Diesel engines. Such engines are best operated at nearly constant speed. Merchant ships do not have to vary their speed much during their voyage; their propeller is diven directly from the engine. Various arrangements are used to allow research vessels to operate at very slow speeds. In diesel-electric systems the Diesel engine powers an electic motor which in turn drives the propeller shaft. Electric motors work efficiently at any speed, allowing the ship very accurate speed control. In another arrangement the Diesel engine drives a "variable pitch propeller". In such a propeller the pitch of the blades (the blade angle) can be controlled to give very low or zero propeller thrust even at full propeller rotation.

Lowering equipment over the side of a vessel requires more than zero thrust. Without active position control the ship will drift with the wind and across the instrument wire. To keep the wire vertical and free from the ship's hull the ship has to counteract the effects of wind and current. This is usually achieved through a pair of additional thrusters, one at the bow and one at the stern, which can push the ship sideways. The bow thruster is either a propeller set into a horizontal tunnel through the ship's hull or a propeller on a shaft that can be oriented in any direction and retracted when the ship is under way. The stern thruster is either a similar tunnel or a propeller built into the ship's rudder (an "active rudder"). The two thrusters together allow very accurate control over the ship's behaviour in wind, waves and current and enable it to turn itself around on the spot if neccessary.

The minimum laboratory requirements consist of a wet laboratory for the handling of water samples, a computer laboratory for data processing, an electronics laboratory for the preparation of instruments and a chemical laboratory for water sample analysis. Larger research vessels designed for multidisciplinary research have additional biological, geophysical and geological laboratories. Figure 13.3 shows a typical deck arrangement on a medium sized research vessel.

Research vessels are expensive to operate (US$15,000 - US$25,000 per day at sea). For many decades they were the only available type of platform for data collection on the high seas. The advent of deep-sea moorings, satellites and autonomous drifters has reduced their importance, but research vessels still are an essential tool in oceanographic research. They are now used principally for large scale near-synoptic surveys of oceanic property fields and for targeted process studies (such as mixing across fronts, determination of the heat budget of small ocean regions etc).

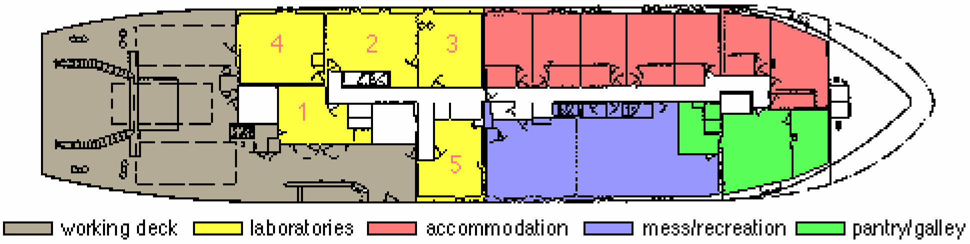


Figure 13.3. Plan of the main deck of the Australian research vessel *Franklin*, showing the main scientist accommodation and laboratory arrangement:

1. wet laboratory (CTD and multiple water sampler area)
2. operations room (echo sounder, ADCP, CTD, towed vehicle and thermosalinograph controls and displays)
3. computer room (terminals for data processing and analysis)
4. chemical laboratory (salinity, nutrient and oxygen analysis)
5. electronic workshop (instrument preparation)

An additional laboratory for chemical or biological work is located on a lower deck. One container laboratory can be placed on the work deck and connected with the chemical laboratory through a water-tight doorway.
Crew accommodation is on the lower and upper decks; the mess room and recreation area is a common area.

Moorings

Moorings are appropriate platforms wherever measurements are required at one location over an extended time period. Their design depends on the water depth and on the type of instrumentation for which the mooring is deployed, but the basic elements of an oceanographic mooring are an anchor, a mooring line (wire or rope) and one or more buoyancy elements which hold the mooring upright and preferably as close to vertical as possible.

Subsurface moorings are used in deep water in situations where information about the surface layer is not essential to the experiment. The main buoyancy at the top of the mooring line is placed some 20 - 50 m below the ocean surface. This has the advantage that the mooring is not exposed to the action of surface waves and is not at risk of being damaged by ship traffic or being vandalised or stolen. Figure 13.4 shows a typical deep-sea mooring. The main buoyancy is at the top of the mooring line. To protect the mooring against fish bites, wire is used for the upper 1000 m or so of the mooring line, while rope is used below.

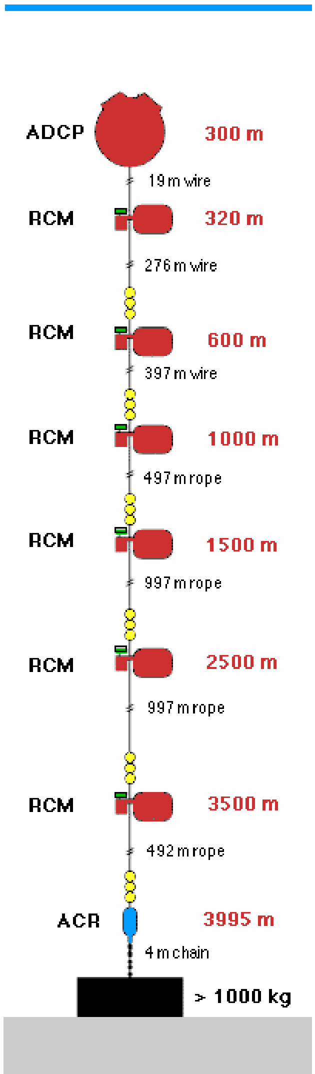


Figure 13.4. A typical deep-sea mooring design sketch for a mooring deployment in 4000 m water depth. Instrument design depths are shown in dark red.

The acoustic doppler current profiler (ADCP) is mounted in the main buoyancy body at 300 m. It looks upward and covers the upper 250 - 300 m of the water column with a resolution of one reading every 8 m.

The first recording current meter (RCM) is placed at 320 m, followed by more recording current meters at regular intervals. Additional buoyancy, shown as yellow clusters of small buoys, is placed above each RCM to compensate its weight in water.

Above the anchor weight is an acoustic release (ACR) which, when triggered, severs the connection to the anchor weight so that the mooring floats to the surface.

To keep the mooring close to the vertical it should have minimum drag, which can only be achieved by keeping the wire diameter small. This requires a small wire load from the instruments. Additional buoyancy is therefore distributed along the wire to compensate for the weight of the instrumentation. The buoyancy is arranged so that all sections of thew mooring are buoyant, enabling recovery of a damaged mooring which lost its upper part.

At the bottom of a deep sea mooring just above the anchor is a remotely controllable release. It can be activated through a coded acoustic signal from the ship when it is time to recover the mooring. Triggering the release brings the mooring to the surface. The anchor, usually a concrete block or a clump of disused railway wheels, is left at the ocean floor.

An experiment which includes the surface layer or the collection of meteorological data requires a **surface mooring**. The main buoyancy for such a mooring takes the shape of a substantial buoy that floats at the surface and can carry meteorological instrumentation (Figure 13.5). In the deep ocean surface moorings are mostly "taut moorings". They use only rope for the mooring line and make it a few per cent shorter than the water depth. This stretches the rope and keeps it under tension to keep the mooring close to vertical. The "inverse catenary" mooring is also used; this is an arrangement where a buoyant section of mooring line is included between two non-buoyant sections causing the profile of the line to form an S-shape. In this configuration the length of the mooring line is not critical and is about 25% greater than the water depth.

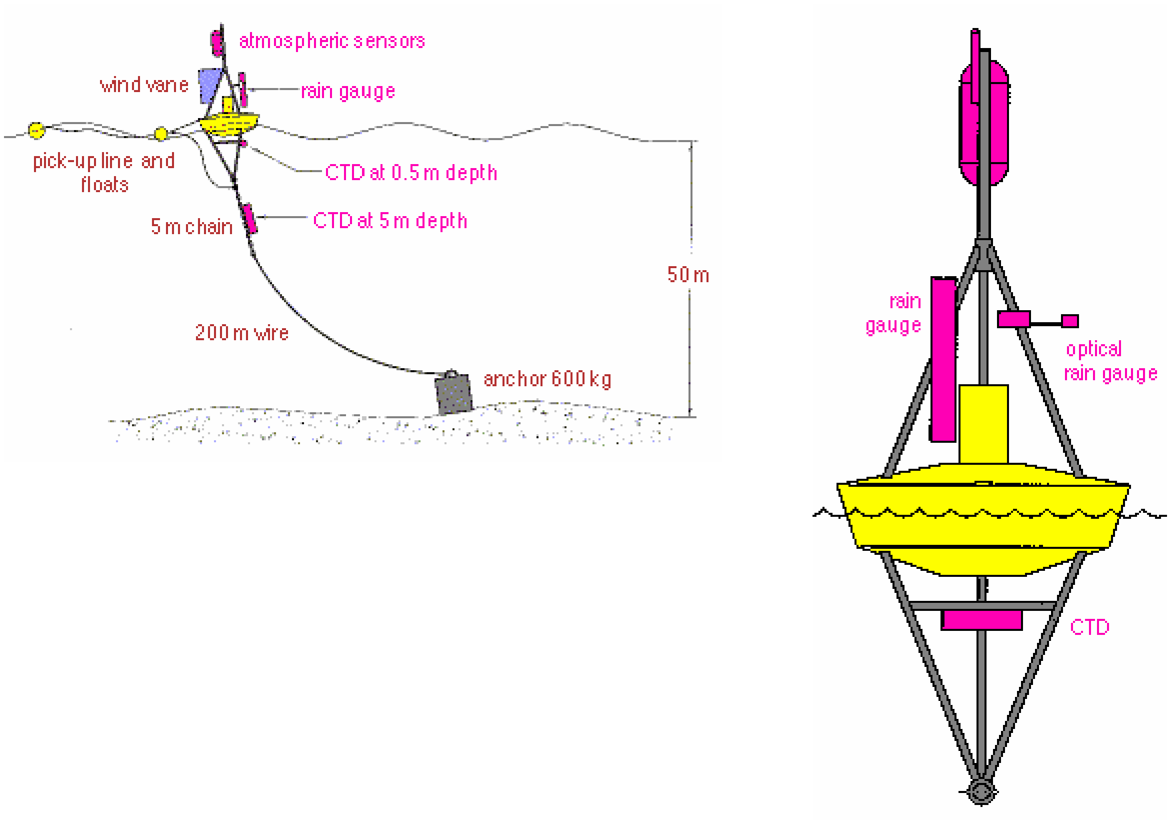


Figure 13.5. A design example for a meteorological surface mooring. This particular mooring was designed for rain measurements, so it carries a wind vane which turns the rain gauge always into the wind, eliminating interference from the buoy's superstructure. Two self-recording CTD systems are attached to the mooring to measure changes of salinity and temperature produced by the rainfall.
This particular design was for quite shallow water depth, so the mooring is a slack mooring.

Moorings on the continental shelf, where the water depth does not exceed 200 m, do not require acoustic releases if a **U-type mooring** is used. A U-type mooring consists of a surface or subsurface mooring to carry the instrumentation, a ground line of roughly twice the water depth and a second mooring with a small marker buoy (Figure 13.6). When the time comes to bring the mooring in, the marker buoy is recovered first, followed by the two anchors and finally the mooring itself. U-type moorings are usually "slack moorings"; the mooring line is longer than the water depth, and the mooring swings with the current.

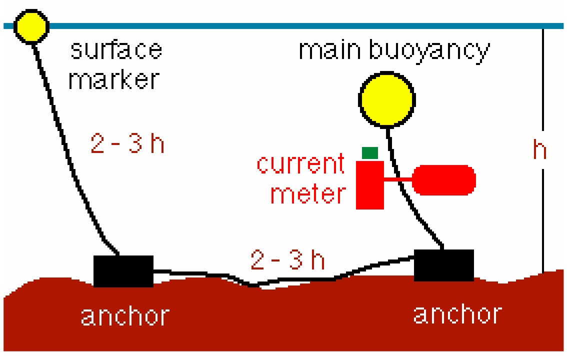


Figure 13.6. A sketch of a U-type mooring. This mooring uses a subsurface buoy to carry the instrumentation (in this example a single current meter; there can of course be more than one instrument on the wire). If the water depth is *h*, the length of the ground line is 2 - 3 *h*, and the slack line below the surface marker is of similar length.

Satellites

The advent of satellite technology opened the possibility of measuring some property fields and dynamic quantities from space. The advantage of this method is the nearly synoptic coverage of entire oceans and ease of access to remote ocean regions. Satellites have therefore become an indispensable tool for climate research. The major restriction of the method is that satellites can only see the surface of the ocean and therefore give only limited information about the ocean interior.

Most satellites are named for the sensors they carry. Strictly speaking, the satellite and its sensors are two separate things; the satellite is a platform, the sensors are instruments. An overview of the available satellite sensors is therefore given in the discussion of instruments below.

As platforms, satellites fall into three groups. Most satellites follow **inclined orbits**: Their elliptical orbits are inclined against the equator. The degree of inclination determines how far away from the equator the satellite can see the Earth. Typical inclinations are close to 60°, so the satellite covers the region from 60°N to 60°S. It covers this region frequently, completing one orbit around the Earth in about 50 minutes.

Some satellites have an inclination of nearly or exactly 90° and can therefore see both poles; they fly on **polar orbits**. A typical height of satellites on polar and on inclined orbits is 800 km.

The third and last group are the **geostationary** satellites. They orbit the Earth at the same speed with which the Earth rotates around its axis and are therefore stationary with respect to the Earth. This situation is only possible if the satellite is over the equator and orbits at a height of 35,800 km, much higher than all other satellites. Geostationary satellites therefore cannot see the poles.

The selection of a satellite as a platform logically includes the selection of a sensor and a suitable orbit. An ice sensor to monitor the polar ice caps does not achieve much on a geostationary satellite; a cloud imager for weather forecasting is not placed in a polar orbit.

Submersibles

Submersibles are not a frequently used platform in physical oceanography, but this is likely to change over the coming years. Three basic types can be distinguished, manned submersibles, remotely controlled submersibles and autonomous submersibles.

Manned submersibles are used in marine geology for the exploration of the sea floor and occasionally in marine biology to study sea floor ecosystems. They are not a tool for physical oceanography.

Remotely controlled submersibles are commonly used in the offshore oil and gas industry and for retrieving flight recorders from aircraft that fell into the ocean. In science they find similar uses to manned submersibles but are again not a tool for physical oceanography.

Autonomous submersibles are self-propelled vehicles that can be programmed to follow a predetermined diving path. Such vehicles have great potential for physical oceanography. Some major oceanographic research institutions are developing vehicles to carry instrumentation such as a CTD and survey an ocean area by regularly diving and surfacing along a track from one side of an ocean region to the other and transmitting the collected data via satellite when at the surface. Some time will pass, however, before these vehicles will come into regular use. Eventually, autonomous submersibles will greatly reduce the need for research vessels for ocean monitoring.

Towed vehicles

Towed vehicles are used from research vessels to study oceanic processes which require high spatial resolution such as mixing in fronts and processes in the highly variable upper ocean. Most systems consist of a hydrodynamically shaped underwater body, an electro-mechanical (often multi-conductor) towing cable and a winch. The underwater body is fitted with a pair of wing shaped fins which control its flight through the water. In addition to the sensor package (usually a CTD, sometimes additional sensors for chemical measurements) it carries sensors for pressure, pitch and roll to monitor its behaviour and control its flight. The data are sent to the ship's computer system via the cable. The same cable is used to send commands to the underwater body to alter its wing angle.

Figure 13.7 shows a towed vehicle during deployment. A typical flight path for this vehicle covers a depth range of about 250 - 500 m, which can be chosen to be anywhere between the surface and 800 m depth. The vehicle is towed at about 6 - 10 knots (10 - 18 km/h) and traverses the 250 m depth range about once every 5 minutes. When fitted with a CTD this results in a vertical section of temperature and salinity with a horizontal resolution of about 1 km.

An alternative towed system does not employ an underwater body to carry the sensor package but has sensors (for example thermistors) built into the towed cable at fixed intervals. Because the distance between the sensors is fixed and the sensors remain at the same depth during the tow, these "thermistor chains" do not offer the same vertical data resolution as undulating towed systems and are only rarely used now.

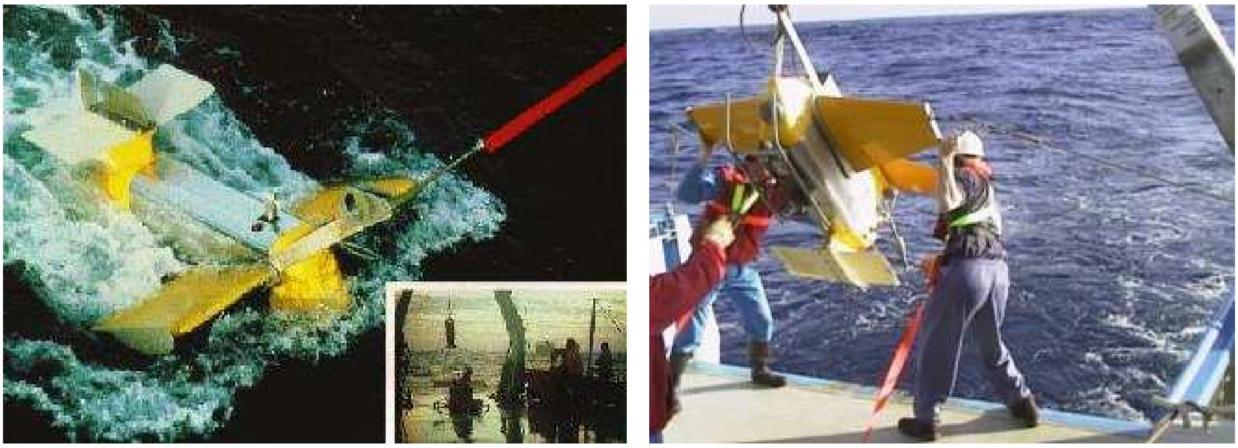


Figure 13.7. Left: "Seasoar", a towed vehicle during deployment. The inset shows the vehicle hanging from the A-frame of the research vessel. In the main figure the vehicle is seen to enter the water. The wings are rotated upward to make it dive. Tail fins give dynamic stability. The instrumentation (a CTD) is housed in the main body. Rigth: A towed vehicle is launched from the Australian research vessel *Franklin*.

Floats and drifters

The main characteristic of floats and drifters is that they move freely with the ocean current, so their position at any given time can only be controlled in a very limited way. Until a decade ago these platforms were mainly used in remote regions such as the Southern Ocean and in the central parts of the large ocean basins that are rarely reached by research vessels and where it is difficult and expensive to deploy a mooring. They have now become the backbone of a new observing system that covers the entire ocean.

Strictly speaking, a float is a generic term for anything that does not sink to the ocean floor. A drifter, on the other hand, is a platform designed to move with the ocean current. To achieve this it has to incorporate a floatation device or float, but it is usually more than that. But oceanographers use the terms very loosely and do not make a clear distinction between "floats" and "drifters".

Two basic types can be distinguished. **Surface drifters** have a float at the surface and can therefore transmit data via satellite. If they are designed to collect information about the ocean surface they carry meteorological instruments on top of the float and a temperature and occasionally a salinity sensor underneath the float. To prevent them from being blown out of the area of interest by strong winds they are fitted with a "sea anchor" at some depth (Figure 13.8). If they are designed to give information on subsurface ocean properties, additional sensors are placed between the surface float and the sea anchor. The depth range of surface drifters is usually limited to less than 100 m.

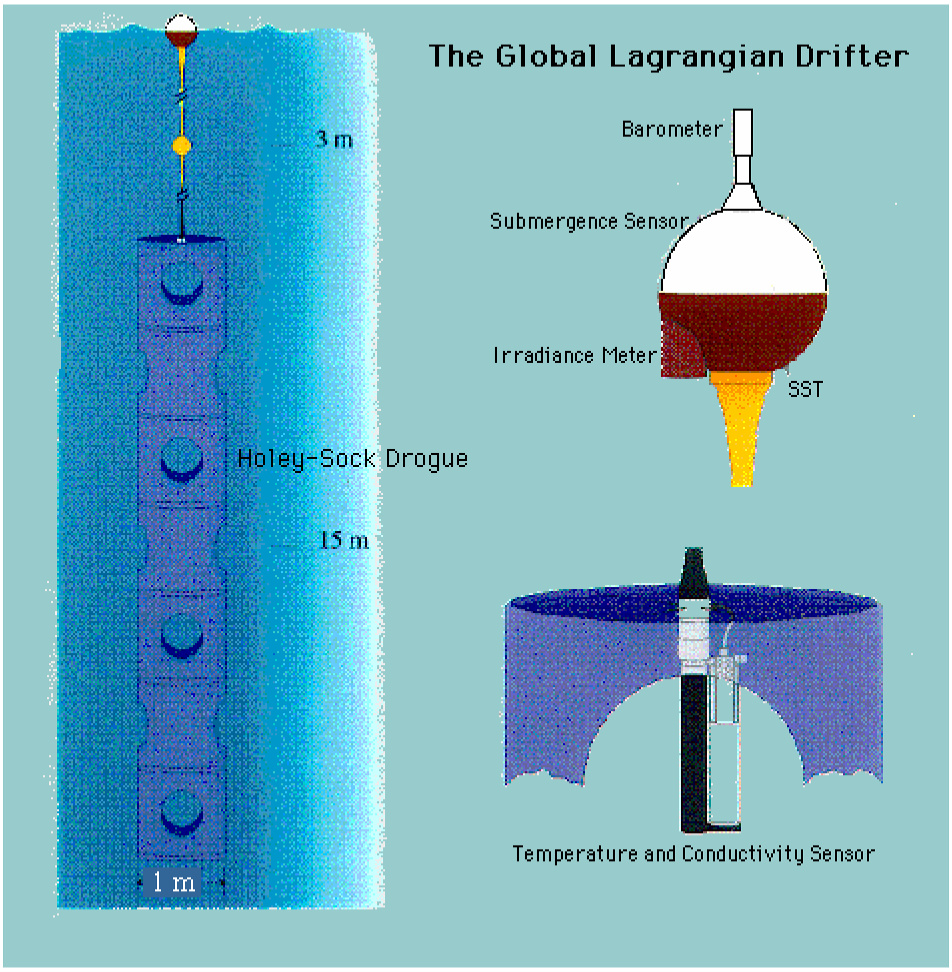


Figure 13.8. A sketch of a modern ocean drifter. The surface float has instruments to measure air pressure, sun light level and sea surface temperature (SST) and a sensor to indicate whether the float is pushed under by strong waves. The sea anchor is a 15 m long tube of 1 m diameter with a series of holes (the "holey-sock drogue"). At the top of the drogue is a mini-CTD. The data are transmitted to shore via satellite.

Floats used for **subsurface drifters** are designed to be neutrally buoyant at a selected depth. These drifters have been used to follow ocean currents at various depths, from a few hundred metres to below 1000 m depth. The first such floats transmitted their data acoustically through the ocean to coastal receiving stations. Because sound travels well at the depth of the sound velocity minimum (the SOFAR channel), these SOFAR floats can only be used at about 1000 m depth.

Modern subsurface floats remain at depth for aperiod of time, come to the surface briefly to transmit their data to a satellite and return to their allocated depth. These floats can therefore be programmed for any depth and can also obtain temperature and salinity (CTD) data during their ascent. The most comprehensive array of such floats, known as Argo, began in the year 2000. Argo floats measure the temperature and salinity of the upper 2000 m of the ocean (Figure 13.8a). This will allow continuous monitoring of the climate state of the ocean, with all data being relayed and made publicly available within hours after collection. When the Argo programme is fully operational it will have 3,000 floats in the world ocean at any one time.

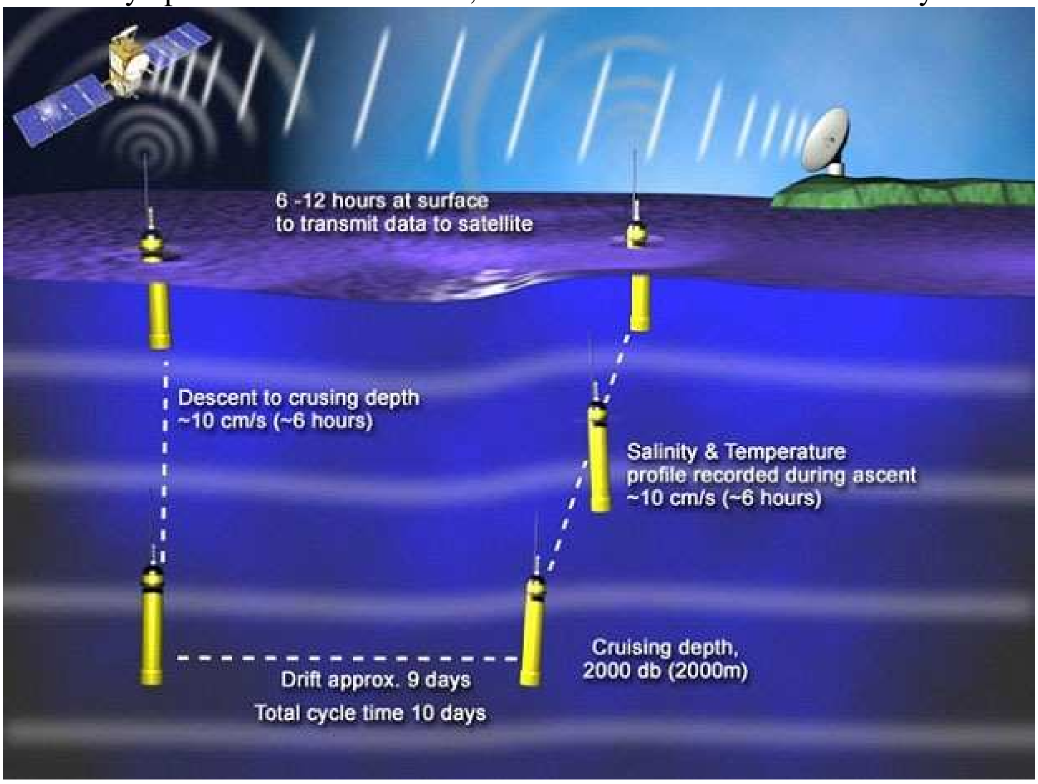


Fig. 13.8a. Principle of operation of an Argo float.

Measurements of hydrographic properties

This section gives an overview of sensors and instrument packages for the measurement of temperature, salinity, oxygen, nutrients and tracers.

Reversing thermometers

The earliest temperature measurements at some depth below the surface where made by bringing a water sample up to the deck of a ship in an insulated bucket and measuring the sample temperature with a mercury thermometer. Although these measurements were not accurate, they gave the first evidence that below the top 1000 m the ocean is cold even in the tropics. They also showed that highly accurate measurements are required to resolve the small temperature differences between different ocean regions at those depths.

The first instrument that (through the use of multiple sampling and averaging) achieved the required accuracy of 0.001°C was the reversing thermometer. It consists of a mercury filled glas pipe with a 360° coil. The pipe is restricted to capillary width in the coil, where it has a capillary appendix (Figure 13.9). The instrument is lowered to the desired depth. Mercury from a reservoir at the bottom rises in proportion to the outside temperature. When the desired depth is reached the thermometer is turned upside down (reversed), but the flow of mercury is now interrupted at the capillary appendix, and only the mercury that was above the break point is collected in the lower part of the glass pipe. This part carries a calibrated gradation that allows the temperature to be read when the thermometer is returned to the surface.

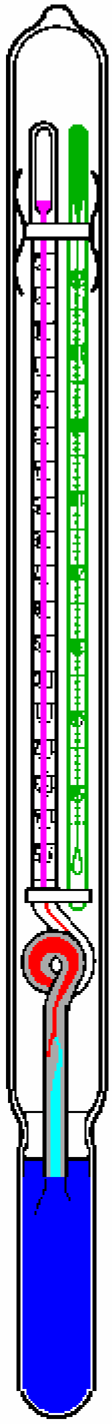


Figure 13.9. A reversing thermometer in the upright (not reversed) position. The mercury rises from the mercury reservoir (blue) through the capillary section (cyan and red) into the glass cylinder and upper reservoir (magenta).

When the thermometer is turned upside down (reversed), the mercury column ruptures at the appendix in the capillary section (between the red and blue). The red and magenta mercury settles into the upper reservoir, while the cyan mercury returns to the blue reservoir. The level of the magenta/red mercury can then be read against the scale and indicates the temperature.

The green thermometer is an auxiliary thermometer which gives the ambient temperature. It is used to check the temperature when the main thermometer is read on deck of the ship. Its reading allows to correct for the expansion of the glass tube that occurs due to the warmer ambient temperature on deck. The overall accuracy thus achieved is 0.02°C. If multiple reversing thermometers are used in parallel the accuracy can be increased to 0.01°C.

To eliminate the effect of pressure, which compresses the pipe and causes more mercury to rise above the break point during the lowering of the instrument, the thermometer is enclosed in a pressure resistent glass housing. If such a "protected reversing thermometer" is used in conjunction with an "unprotected reversing thermometer" (a thermometer exposed to the effect of pressure), the difference between the two temperature readings can be used to determine the pressure and thus the depth at which the readings were taken. The reversing thermometer is thus also an instrument to measure depth.

Reversing thermometers require a research vessel as platform and are used in conjunction with Nansen or Niskin bottles or on multi-sample devices.

Nansen and Niskin bottles

The measurement of salinity and oxygen, nutrients and tracer concentrations requires the collection of water samples from various depths. This essential task is achieved through the use of "water bottles". The first water bottle was developed by Fritjof Nansen and is thus known as the Nansen bottle. It consists of a metal cylinder with two rotating closing mechanisms at both ends. The bottle is attached to a wire (Figure 13.10). When the bottle is lowered to the desired depth it is open at both ends, so the water flows through it freely. At the depth where the water sample is to be taken the upper end of the bottle disconnects from the wire and the bottle is turned upside down. This closes the end valves and traps the sample, which can then be brought to the surface.

Figure 13.10. The Nansen bottle, invented in 1910, has been widely replaced by the Niskin bottle and is no longer manufactured commercially. The photo on the left is taken from a report of Fridtjof Nansen's work. It shows a Nansen bottle being prepared for sampling. The top of the bottle (not visible) is already clamped to the wire; the scientist attaches the bottom clamp of the bottle.
When the target depth is reached a metal weight (the "messenger") comes down the wire and releases the clamp at the top. The bottle turns over and now hangs upside down from the bottom clamp. This action closes the bottle. (The rod between the bottle and the wire is part of the closing mechanism for the top end.)
A frame at the front of the bottle (not very clearly visible between bottle and scientist in the photo) holds up to three reversing thermometers. They turn over (reverse) with the bottle.
Below the bottle on the wire is another messenger. It is attached to the bottom clamp and released when the bottle turns over. It then travels down the wire and operates another Nansen bottle further down in the water column.

In an "oceanographic cast" several bottles are attached at intervals on a thin wire and lowered into the sea. When the bottles have reached the desired depth, a metal weight ("messenger") is dropped down the wire to trigger the turning mechanism of the uppermost bottle. The same mechanism releases a new messenger from the bottle; that messenger now travels down the wire to release the second bottle, and so on until the last bottle is reached.

The Nansen bottle has now widely been displaced by the Niskin bottle (Figure 13.11). Based on Nansen's idea, it incorporates two major modifications. Its cylinder is made from plastic, which eliminates chemical reaction between the bottle and the sample that may interfere with the measurement of tracers. Its closing mechanism no longer requires a turning over of the bottle;

the top and bottom valves are held open by strings and closed by an elastic band. Because the Niskin bottle is fixed on the wire at two points instead of one (as is the case with the Nansen bottle) it makes it easier to increase its sample volume. Niskin bottles of different sizes are used for sample collection for various tracers.

Figure 13.11. A scientist draws a water sample from a Niskin bottle mounted on a multiple water sample device. The long white cylinder is the Niskin bottle. The white plastic frame attached to the Niskin bottle can hold up to three reversing thermometers. It rotates around its centre when the Niskin bottle closes. (Photo © 2000 CSIRO Marine Research, reproduced by permission.) Figure 13.13 gives another view of Niskin bottles on a multiple water sample device.

The figure below shows a Niskin bottle designed for use on a wire. The bottle is closed and on its side; the thermometer frame is at the bottom. The bottle is attached to the wire with the two wing nuts at the top.

Nansen and Niskin bottles are used on conjunction with reversing thermometers. On the Nansen bottle the thermometers are mounted in a fixed frame, the reversal being achieved by the turning over of the bottle. On Niskin bottles thermometers are mounted on a rotating frame.

CTDs

Today's standard instrument for measuring temperature, salinity and often also oxygen content is the CTD, which stands for **c**onductivity, **t**emperature, **d**epth (Figure 13.12). It employs the principle of electrical measurement. A platinum thermometer changes its electrical resistance with temperature. If it is incorporated in an electrical oscillator, a change in its resistance produces a change of the oscillator frequency, which can be measured. The conductivity of seawater can be measured in a similar way as a frequency change of a second oscillator, and a pressure change produces a frequency change in a third oscillator. The combined signal is sent up through the single conductor cable on which the CTD is lowered. This produces a continuous reading of temperature and conductivity as functions of depth at a rate of up to 30 samples per second, a vast improvement over the 12 data points produced by the 12 Nansen or Niskin bottles that could be used on a single cast.

Figure 13.12. Two examples of CTD instruments. The CTD on the left is designed for deep ocean measurements. Its sensors for temperature, conductivity and pressure are contained in the small packages at the bottom corners of the frame. The main cylinder houses the control and data processing electronics. The entire package is lowered on a conducting cable and connected through the sets of electrical underwater connectors at the top.
The instrument at the right is a lighter version of the same principle for use on small boats in estuaries and coastal waters. The sensors are seen at the left. The instrument is lowered on a rope, without electrical connection with the boat. When it is brought back on deck it is connected to a computer, and the data are downloaded. The computer screen shows a typical result of a measurement station: Temperature (white), conductivity (green), salinity (blue) and density (red) against depth.

Electrical circuits allow measurements in quick succession but suffer from "instrumental drift", which means that their calibration changes with time. CTD systems therefore have to be calibrated by comparing their readings regularly against more stable instruments. They are therefore always used in conjunction with reversing thermometers and a multi-sample device.

Multiple water sample devices

Multiple water sample devices allow the use of Niskin bottles on electrically conducting wire. Different manufacturers have different names for their products, such as *rosette* or *carousel*. In all products the Niskin bottles are arranged on a circular frame (Figure 13.13), with a CTD usually mounted underneath or in the centre.

Figure 13.13. Two examples of a multiple water sample device with Niskin bottles. The device in the first figure is a design for shallow water and estuarine applications; it carries 12 Niskin bottles of relatively small volume. Two bottles are shown in the open position, when the water can flow through. The remaining bottles are shown in the closed position. The second figure shows a system for deep ocean use with 24 Niskin bottles and thermometer holders for reversing thermometers. Some bottles on the left are open, with the thermometer holder in the upright position. All bottles at the front are closed, and their thermometer holders are in the reversed position.

The advantage of multi-sample devices over the use of a hydrographic wire with messengers is that the water bottles can be closed by remote control. This means that the sample depths do not have to be set before the bottles are lowered. As the device is lowered and data are received fom the CTD, the operator can look for layers of particular interest and take water samples at the most interesting depth levels.

Thermosalinographs

The introduction of the CTD opened the possibility of taking continuous readings of temperature and salinity at the surface. Water from the cooling water intake of the ship's engines is pumped through a tank in which a temperature and a conductivity sensor are installed. Such a system is called a thermosalinograph.

Remote sensors

Most oceanographic measurements from space or aircraft are based on the use of radiometers, instruments that measure the electro-magnetic energy radiating from a surface. This radiation occurs over a wide range of wavelengths, including the emission of light in the visible range, of heat in the infrared range, and at shorter wavelengths such as Radar and X-rays. Most oceanographic radiometers operate in several wavelength bands. A detailed discussion of all applications goes beyond the scope of these lecture notes; so only the most basic systems are mentioned here.

Radiometers that operate in the infrared are used to measure sea surface temperature. Their resolution has steadily increased over the years; the **AVHRR** (**A**dvanced **V**ery **H**igh **R**esolution **R**adiometer) has a resolution that comes close to 0.2°C.

Multi-spectral radiometers measure in several wavelength bands. By comparing the radiation signal received at different wavelengths it is possible to measure ice coverage and ice age, chlorophyll content, sediment load, particulate matter and other quantities of interest to marine biology.

Measurements at radar wavelengths are made by an instrument known as **SAR** (**S**ynthetic **A**perture **R**adar). It can be used to detect surface expressions of internal waves, the effect of rainfall on surface waves, the effect of bottom topography on currents and waves, and a range of other phenomena. Many of these phenomena belong into the category "dynamic properties" discussed below.

Measurements of dynamic properties

All instruments discussed so far produce information about oceanic property fields irrespective of the dynamic state of the ocean. The remainder of this lecture summarises instrumentation designed to measure movement in the ocean.

An elementary way of observing oceanic movement is the use of drifters. As mentioned above, drifters are platforms designed to carry instruments. But all measurements obtained from drifters are of little use unless they can be related to positions in space. A geolocation (GPS) device which transmits the drifter location to a satellite link is therefore an essential instrument on any drifter, and this turns the drifter into an instrument for the measurement of ocean currents. Whether it does that job well or not depends on its design, and in particular the size and shape of its sea anchor.

Current meters

Ocean currents can be measured in two ways. An instrument can record the speed and direction of the current, or it can record the east-west and north-south components of the current (Figure 13.14). Both methods require directional information. All current meters therefore incorporate a magnetic compass to determine the orientation of the instrument with respect to magnetic north. Four classes of current meters can be distinguished, based on the method used for measuring current magnitude.

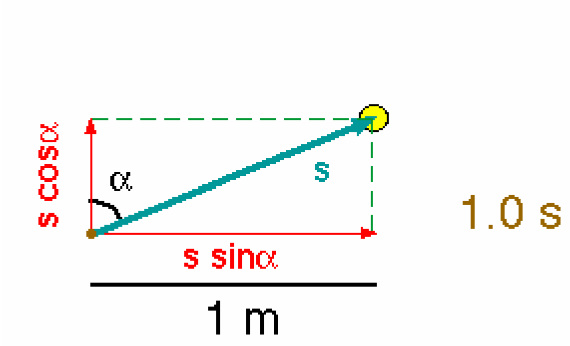


Figure 13.14. A sketch of the two ways in which movement can be quantified. The movement is indicated by the yellow ball and shown in time steps of 0.2 seconds. It indicates steady movement towards α = 68.2° east of north. (Note that direction is expressed in degrees from North and clockwise.) After one second the ball has moved the distance s = 1.077 m. This movement can therefore be described by the two numbers
speed *s*: 1.077 m s^{-1}
direction α: 68.2°

An alternative way of describing the movement is to give its components along the two axes of the co-ordinate system (shown by the red arrows). The east component is then given by $s \cdot \sin\alpha$, the north component by $s \cdot \cos\alpha$. In this example the two components are
east component: 1 m s^{-1}
north component: 0.4 m s^{-1}

Mechanical current meters use a propeller-type device, a Savonius rotor or a paddle-wheel rotor (Figure 13.15) to measure the current speed, and a vane to determine current direction. Propeller sensors often measures speed correctly only if they point into the current and have to be oriented to face the current all the time. Such instruments are therefore fitted with a large vane, which turns the entire instrument and with it the propeller into the current.

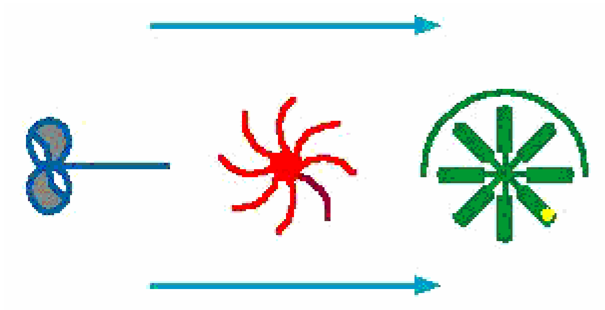


Figure 13.15. The three mechanical methods of measuring currents. The example assumes that the instruments are exposed to a current which flows from west to east, then pauses, then flows from east to west with the same strength as before, so that there is no net flow (the mean current is zero). It is also assumed that the orientation of the speed sensors is not changed in space during the experiment.
The propeller rotor (left) turns as the current flows from west to east, stops, and turns backward as the current flows from east to west. But the backward turns do not cancel the forward turns because the propeller response to flow from its back is different to its response to flow coming from the front (a result of flow obstruction from the axle and instrument housing located behind the axle). Unless the propeller is specifically designed to have a cosine response to the flow, a propeller current meter thus only gives a reliable reading if the propeller always points into the current and its orientation is recorded along with the measured current speed.
The Savonius rotor (centre) turns anti-clockwise, stops, and continues to turn anticlockwise, effectively integrating over the current speed. It therefore gives a large apparent current speed. It does this independently of its orientation in space. A Savonius rotor thus gives a reliable reading only in situations where there is no high frequency alternating flow (such as produced, for example, close to the ocean surface by wind waves). Its reading has to be supplemented by independent information on current direction.
The paddle-wheel rotor (right) does not rotate at all unless one side of it is sheltered from the current. If it is sheltered, the paddle-wheel turns anti-clockwise first, then stops, then turns clockwise, and the clockwise rotation cancels the anti-clockwise rotation exactly. A paddle-wheel rotor thus produces a reliable reading. Its reading has to be supplemented by independent information on current direction.

Propellers can be designed to have a cosine response with the angle of incidence of the flow. Two such propellers arranged at 90° will resolve current vectors and do not require an orienting vane.

The advantage of the Savonius rotor is that its rotation rate is independent of the direction of exposure to the current. A Savonius rotor current meter therefore does not have to face the current in any particular way, and its vane can rotate independently and be quite small, just large enough to follow the current direction reliably.

With the exception of the current meter that uses two propellers with cosine response set at 90° to each other, mechanical current meters measure current speed by counting propeller or rotor revolutions per unit time and current direction by determining the vane orientation at fixed intervals. In other words, these current meters combine a time integral or mean speed over a set

time interval (the number of revolutions between recordings) with an instanteneous reading of current direction (the vane orientation at the time of recording). This gives only a reliable recording of the ocean current if the current changes slowly in time. Such mechanical current meters are therefore not suitable for current measurement in the oceanic surface layer where most of the oceanic movement is due to waves.
The Savonius rotor is particularly problematic in this regard. Suppose that the current meter is in a situation where the only water movement is from waves. The current then alternates back and forth, but the mean current is zero. A Savonius rotor will pick up the wave current irrespective of its direction, and the rotation count will give the impression of a strong mean current. The paddle-wheel rotor is designed to rectify this; the paddle wheel rotates back and forth with the wave current, so that its count represents the true mean current (Figure 13.16).

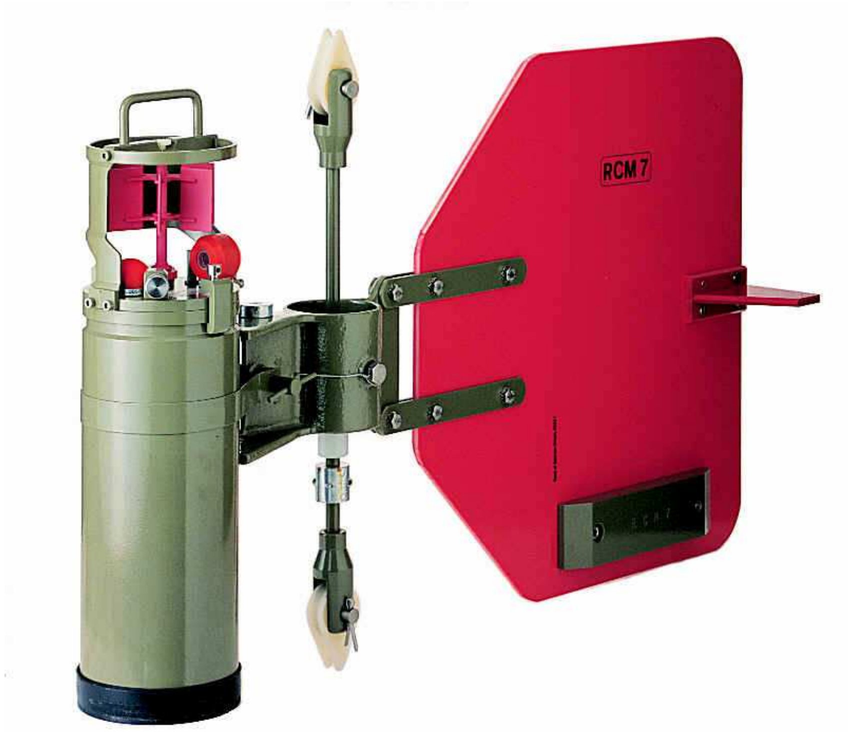


Figure 13.16. A mechanical current meter with a paddle-wheel rotor. When the instrument is placed in a mooring the wire connects to the steel rod at the top and at the bottom. The instrument rotates freely around the rod and is pointed into the current by the big vane at the right. The paddle wheel is shielded on one side, so it turns when the current goes past. Below the paddle wheel are sensors for temperature, conductivity and pressure.

Mechanical current meters are robust, reliable and comparatively low in cost. They are therefore widely used where conditions are suitable, for example at depths out of reach of surface waves.

Electromagnetic current meters exploit the fact that an electrical conductor moving through a magnetic field induces an electrical current. Sea water is a very good conductor (see Lecture 3), and if it is moved between two electrodes the induced electrical current is proportional to the ocean current velocity between the electrodes. An electromagnetic current meter has a coil to produce a magnetic field and two sets of electrodes, set at right angle to each other, and determines the rate at which the water passes between both sets. By combining the two components the instrument determines speed and direction of the ocean current.

Acoustic current meters are based on the principle that sound is a compression wave that travels with the medium. Assume an arrangement with a sound transmitter between two receivers in an ocean current. Let receiver *A* be located upstream from the transmitter, and let receiver *B* located downstream. If a burst of sound is generated at the transmitter it will arrive at receiver *B* earlier than at receiver *A*, having been carried by the ocean current.

A typical acoustic current meter will have two orthogonal sound paths of approximately 100 mm length with a receiver/transmitter at each end. A high frequency sond pulse is transmitted simultaneously from each transducer and the difference in arrival time for the sound travelling in opposite directions gives the water velocity along the path.

Electromagnetic and acoustic current meters have no moving parts and can therefore take measurements at a very high sampling rate (up to tens of readings per second). This makes them useful not only for the measurement of ocean currents but also for wave current and turbulence measurements.

Acoustic doppler current profilers (ADCPs) operate on the same principle as acoustic current meters but have transmitter and receiver in one unit and use reflections of the sound wave from drifting particles for the measurement. Seawater always contains a multitude of small suspended particles and other solid matter that may not all be visible to the naked eye but reflects sound. If sound is transmitted in four inclined beams at right angle to each other, the Doppler frequency shift of the reflected sound gives the reflecting particle velocity along the beam. With at least 3 beams inclined to the vertical the 3 components of flow velocity can be determined. Different arrival times indicate sound reflected at different distances from the transducers, so an ADCP provides information on current speed and direction not just at one point in the ocean but for a certain depth range; in other words, an ADCP produces a current profile over depth.

Different ADCP designs serve different purposes (Figure 13.17). Deep ocean ADCPs have a vertical resolution of typically 8 metres (they produce one current measurement for every 8 metres of depth increase) and a typical range of up to 400 m. ADCPs designed for measurements in shallow water have a resolution of typically 0.5 m and a range of up to 30 m. ADCPs can be placed in moorings, installed in ships for underway measurements, or lowered with a CTD and multi-sample device to give a current profile over a large depth range.

Figure 13.17. Two examples of an ADCP. The orange coloured parts are the sound transmitters/receivers, the white cylinder contains the electronics. The instrument on the left is designed for deep water applications; it has a resolution of 32 m and a range of 600 m. The instrument on the right is a typical shallow water design; its resolution can be varied between 1 m and 4 m, which gives it a range from 20 m to 110 m.

Wave measurements

The parameters of interest in the measurement of surface waves are wave height, wave period and wave direction. At locations near the shore wave height and wave period can be measured by using the principle of the stilling well described for tide gauges below, with an opening wide enough to let pass surface waves unhindered. Wave measurements on the shelf but at some distance from the shore can be obtained from a pressure gauge (see under tide gauges as well).

An instrument suitable for all locations including the open ocean is the **wave rider**, a small surface buoy on a mooring which follows the wave motion. A vertical accelerometer built into the wave rider measures the buoy's acceleration generated by the waves. The data are either stored internally for later retrieval or transmitted to shore. Wave riders provide information on wave height and wave period. If they are fitted with a set of 3 orthogonal accelerometers they also record wave direction.

Tide gauges

Tides are waves of long wavelength and known period, so the major properties of interest for measurement are wave height, or tidal range, and wave induced current. The latter is measured with current meters; any type is suitable. Two types of tide gauges are used to measure the tidal range. The **stilling-well gauge** consists of a cylinder with an connection to the sea at the bottom. This connection acts as a low pass filter: It is so restricted that the backward and forward motion of the water associated with wind waves and other waves of short period cannot pass through; only the slow change of water level associated with the tide can enter the well. This change of water level is picked up by a float and recorded (Figure 13.18).

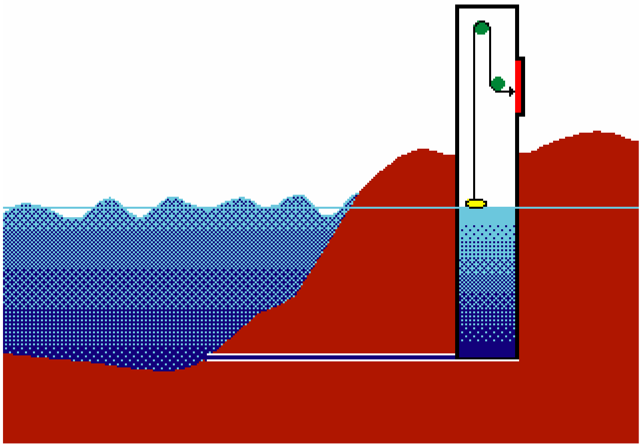


Figure 13.18. Sketch of a stilling-well tide gauge. The gauge is connected to the sea below the level of the lowest tide via a narrow tube. Rapid fluctuations of sea level produced by (for example) wind waves cannot penetrate this tube because its small diameter does not allow rapid transport of water through it.
The thin line indicates the sea level after removal of wind waves and other high frequency fluctuations (the "low-passed sea level"). Slow up and down movement of this sea level enters the still well, so the water level in the well is always the same as the low-passed sea level. A float is connected to a recording pen which writes on a rotating drum (red). This produces a sea level record on a paper attached to the drum.
Modern stilling-well gauges use non-mechanical means of measuring the water level in the well (acoustic or laser) and record the data electronically, but the principle remains the same.

Stilling-well gauges allow the direct reading of the water level at any time but require a somewhat laborious installation and are impracticable away from the shore. In offshore and remote locations it is often easier to use a **pressure gauge**. Such an instrument is placed on the sea floor and measures the pressure of the water column above it, which is proportional to the height of water above it. The data are recorded internally and not accessible until the gauge is recovered.

Tide gauges are increasingly used to monitor possible long term changes in sea level linked with climate variability and climate change. The expected rate of sea level change is only a few millimetres per year at most, so very high accuracy is required to verify such changes. Most tide gauges are not suitable for such a task, for a number of reasons. For example, a long term trend in observed sea level can also be produced by a rise or fall of the land on which the tide gauge is built. (This is known as benchmark drift.) The wire in a stilling-well gauge that connects the float with the recording unit stretches and shrinks as the air temperature rises and falls. Such effects are insignificant when the gauge is used to verify the depth of water for shipping purposes but not when it comes to assessing trends of millimetres per year. A new generation of tide gauges is being installed world wide which gives water level recordings to absolute accuracies of a few millimetres with long term benchmark stability. In these instruments the float and wire arrangement of the stilling-well gauge is replaced by a laser distance measurement, and the data are transmitted via satellite to a world sea level centre which monitors the performance of every gauge continuously.

Remote sensors

Sea level can also be measured from satellites. An **altimeter** measures the distance between the satellite and the sea surface. If the satellite position is accurately known this results in a sea level measurement. Modern altimeters have reached an accuracy of better than 5 cm. The global coverage provided by satellites allows the verification of global tide models. When the tides are subtracted, the measurements give information about the shape of the sea surface and, through application of the principle of geostrophy, the large scale oceanic circulation.

Shear probes

This extremely brief overview of oceanographic measurement techniques can only cover the essentials of the most important platforms and instruments. Special equipment exists, and new special equipment is being designed every day, to address specific problems. The shear probe may serve as an example. It is designed to give insight into oceanic turbulence at the centimetre scale. Turbulence is characterised by currents which vary over short distances and short time intervals, so an instrument designed to measure turbulence has to be able to resolve differences in current speed and direction over a vertical distance of not more than a metre or so.

One such shear probe is a cylindrical instrument of less than 1 m length with two electomagnetic or acoustic current meters at its two ends. By measuring current speed and direction at two points less than 1 m apart it allows the determination of the current shear over that distance. To allow a reliable measurement not influenced by the heaving motion of the ship the probe falls slowly and freely through the ocean. Its maximum diving depth is programmed before the experiment, and the probe returns to the surface when that depth is reached. It is then picked up by the ship, and the internally recorded data are retrieved.

Another type of free-fall instrument uses microstructure sensors that measure velocity fluctuations on a spacial scale of about 10 mm. It uses a piezo-electric beam that generates small voltages as the turbulent velocity varies the lift and thus the bending force on an aerofoil as it moves through the water.