OCEAN CIRCULATION

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Introduction

This article discusses the following aspects of ocean circulation: what is meant by the term ocean circulation; how the ocean circulation is determined by measurements and dynamical processes; the consequences of this circulation on the Earth's climate.

What is Ocean Circulation?

The ocean circulation in its simplest form is the movement of sea water through the ocean, which principally transfers temperature and salinity, from one region to another. Temperature differences between regions cause heat transfers. Similarly, differences in salinity produce transfers of salt. On the time scale of the ocean circulation the inputs and exports of salt into and out of the ocean make a negligible contribution to overall salinity and so variations in salinity occur by the addition and removal of fresh water into and out of the ocean.

Two major processes control the ocean circulation: the action of the wind and the action of small density differences, produced by differences in temperature and salinity, within the ocean. The former process is the wind driven circulation (*see* **Wind Driven Circulation**) whereas the latter is the thermohaline circulation (*see* **Thermohaline Circulation**). Although it is useful to separate these two processes to better understand the ocean circulation, they are not independent from each other.

Ocean circulation is in reality a very complex system, as the flows are not steady in time or space. They are turbulent flows that show variability on scales from the largest scale of the ocean basins (*see* **El Niño Southern Oscillation** (**ENSO**), **North Atlantic Oscillation** (**NAO**)) to the smallest scales where the energy is finally dissipated as heat (*see* **Waves on Beaches**). This turbulent structure of the ocean means there are fundamental limitations on the predictability of its behavior.

Because of this inherent complexity oceanographers have approached ocean circulation by using a combination of observational methods including ships (*see* **Ships**), buoys (*see* **Drifters and Floats**), (*see* **Moorings**) and satellites (*see* **Satellite Altimetry**), combined with the mathematical methods of dynamical oceanography (*see* **Elemental Distribution: Overview, General Circulation Models, Inverse Models**). This integrated approach allows hypotheses to be made that can be tested by comparison with observations. Furthermore, mathematical models of the ocean circulation, based on the dynamical principles, can be constructed and tested against observations (*see* **El Niño Southern Oscillation (ENSO) Models, Regional and Shelf Sea Models**).

This article considers how ocean circulation is measured, how the major processes at work are determined and the consequences of the ocean circulation on the climate system.

How is the Ocean Circulation Determined?

The determination of ocean currents involves measurement of the displacement of an element of fluid over a measured time interval. The position of the measurement is defined mathematically in a Cartesian coordinate system (**Figure 1**) where x is positive eastward direction (lines of constant latitude), y is positive northward direction (geographic North Pole), and z is positive upwards; $z = 0$ corresponds to mean sea level. Without ocean currents and tides the sea level would be an equipotential surface, i.e., one of constant potential energy. The z coordinate is perpendicular to the equipotential surface. The origin is the intersection of the Greenwich meridian (Universal meridian) and the equator with mean sea level.

The coordinates of a parcel of water can be determined by the Global Positioning System (GPS). This satellite-based system provides a horizontal position with an accuracy of better than 100 m, which is sufficient for large-scale flows in the ocean. Large-scale flows are at least 10 km in spatial scale and have timescales of at least a day. A pressure device attached to the current meter normally determines the vertical position.

There are two mathematical methods for defining the displacement of the fluid. One is to measure the velocity of the fluid at a fixed point in the ocean, and the other is to follow the element of the fluid and to measure its velocity as it moves through the ocean. The first method is known as a Eulerian description and the second is a Lagrangian description of flow. In principle, the two methods are independent of each other. This means that a

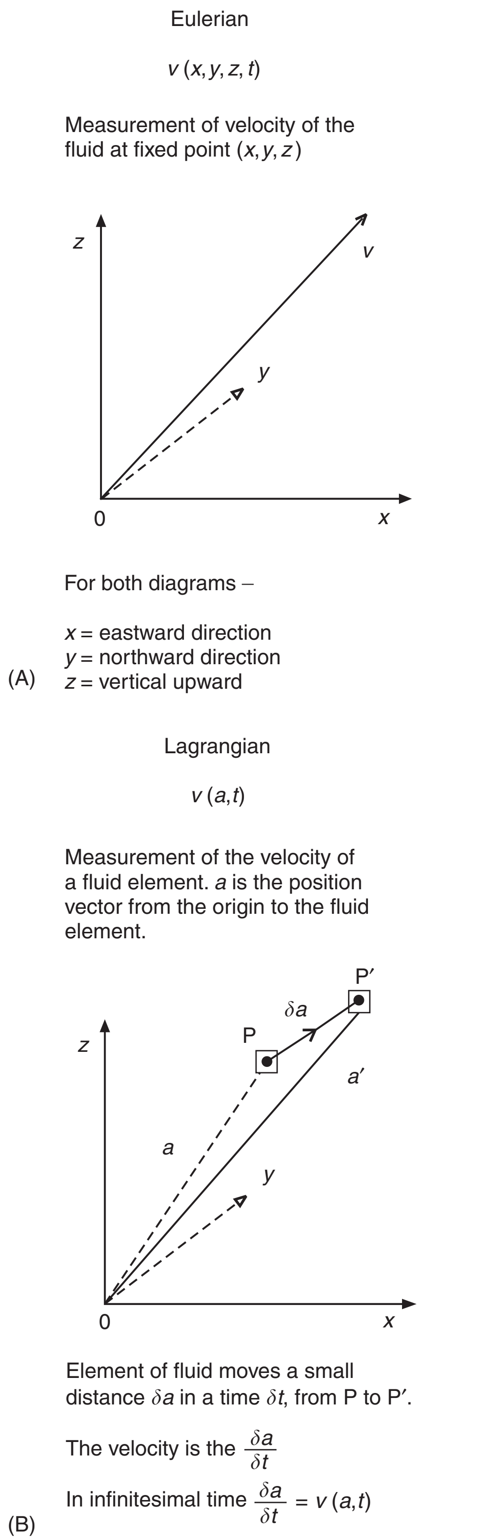

Figure 1 Eulerian and Lagrangian specifications of flow.

Eulerian measurement can not provide Lagrangian currents, and Lagrangian measurements can not provide Eulerian currents.

Having defined the two mathematical methods how the currents are measured in practice is now considered. Initially, these methods will address only the measurement of the horizontal flow. The vertical flow is difficult to measure directly, and will be discussed later in this article.

First, the Eulerian method is considered. The measurement of the flow at a fixed point in the ocean is only straightforward when a fixed position can be maintained, for instance with a current meter attached to the bottom of the ocean or to a pier on the coast. Most measurements have to be made well away from land. This is achieved by attaching the current meter to a mooring (*see* **Moorings**) which is attached to weights and then deployed (**Figure 2**). The position of a mooring can be determined by GPS. The current meter may be a rotary device or an acoustic device. The rotary current meter measures the number of revolutions over a fixed period, whereas the acoustic one measures the change in frequency of an emitted sound pulse caused by the ocean current (i.e., it uses the Doppler effect). Moorings may be deployed for periods of up to 2 years. In the analysis of the record it is normal to remove the high frequency variability of less than 1 day caused by tides by filtering the data.

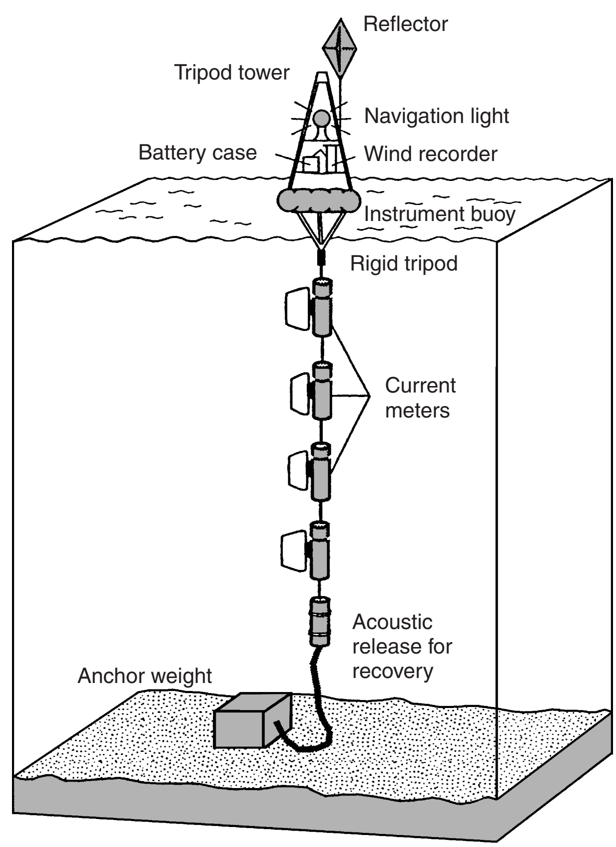

Figure 2 A typical current meter mooring.

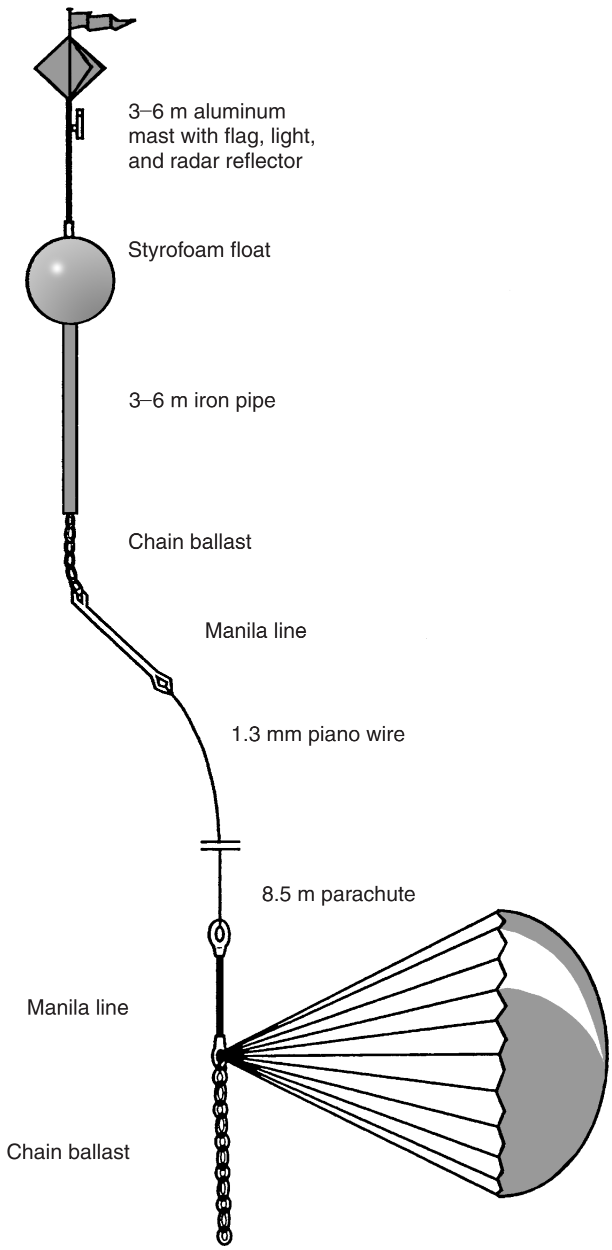

Figure 3 A typical drifter with a parachute drogue of a few meters below the surface. It will follow the current at the depth of the parachute.

A Lagrangian measurement of current can be determined by following an element of water with a float. The horizontal displacement of the water over a small interval of time defines the Lagrangian current. **Figure 3** shows typical float designs that are used (*see* **Drifters and Floats**). The position of the float can be determined by two methods. A float that has a surface satellite transmitter/receiver can have its position determined by GPS, whereas a subsurface float would use an acoustic navigation system. Some floats can descend to a predetermined depth, maintain that depth for a few weeks and then return to the surface for a position fix. This technique allows the current to be measured down to depths of 1 km below the surface.

Each method gives different information on the flow field. A mooring will give a time series of the horizontal current, whereas a float will give the trajectory of the horizontal displacement of the parcel. It is worth remarking that most of the information on the surface ocean circulation has come from mariners' observations of the ships set, a method which has been used since the nineteenth century. However, these measurements have their limitations since they are neither eulerian nor lagrangian measurements and additional influences (e.g., wind effects on the ship) may cause errors.

This information can be analyzed in many different ways to discern the major current systems. From a set of moorings deployed for a few years across, say, the Gulf Stream, the mean flow (i.e., the average of all the current measurements) and its variability can be determined. The mean flow could be calculated over a particular time period. This time period is limited by the period of deployment, which is of the order of 2 years. This is rather short for a climatological mean, and a much longer period of 10 years is desirable. A few longer time series of currents have been determined for the Gulf Stream in the Florida Straits (*see* **Rigs and Offshore Structures**) and for the Antarctic Circumpolar Current in the Drake Passage (*see* **Antarctic Circumpolar Current**).

Recall ocean currents are turbulent and therefore have variability on a whole range of timescales. Hence the mean flow gives no information on the variability of the flow. However, the statistics of the flow can be calculated, based on the kinetic energy (KE). The kinetic energy/unit volume is defined as:

$$\mathrm{KE} = \frac{1}{2}\rho[u^2 + v^2]$$

where ρ is the density of the sea water and u and v are the eastward and northward components of the horizontal flow, respectively.

If the time mean current is defined as $\bar{u}$ and u' as the deviation from $\bar{u}$ at any time, the mean kinetic energy (KEM) and eddy kinetic energy (EKE) can be defined by:

$$\mathrm{KEM} = \frac{1}{2}\rho[\bar{u}^2 + \bar{v}^2]$$

$$\mathrm{EKE} = \frac{1}{2}\rho[u'^2 + v'^2]$$

These two numbers give quantitative measures on the mean and variability of the flow respectively. The ratio EKE/KEM gives a measure of the relative variability of the flow. If the ratio is very much less than 1 then the flow is steady, whereas if the ratio is approximately equal to 1 then the flow is very variable.

Figure 4 shows the variability of the flow in the Agulhas Current (*see* **Agulhas Current**), which is an intense and highly variable current off the coast of South Africa.

Although this ratio gives a measure of the variability of the current, it does not give any idea of the exact time or space scales over which the current is varying. For example, the current may show a slow change from one season to another or it may show faster variation due to eddies.

To address this variation time series analysis, such as Fourier analysis, can be used to determine the KE of the flow for different time periods. Fourier analysis produces a spectrum of the KE, either in frequency for a time series, or in wave number for a spatial variation in flow. **Figure 5** shows the analysis of a time series into its component frequencies. If the current is varying on all timescales the spectrum would be flat, but if there was only one dominant period, it would peak at that one frequency. This particular analysis shows that the current is varying at the tidal frequency and the inertial frequency both at the high frequency end of the spectrum. The

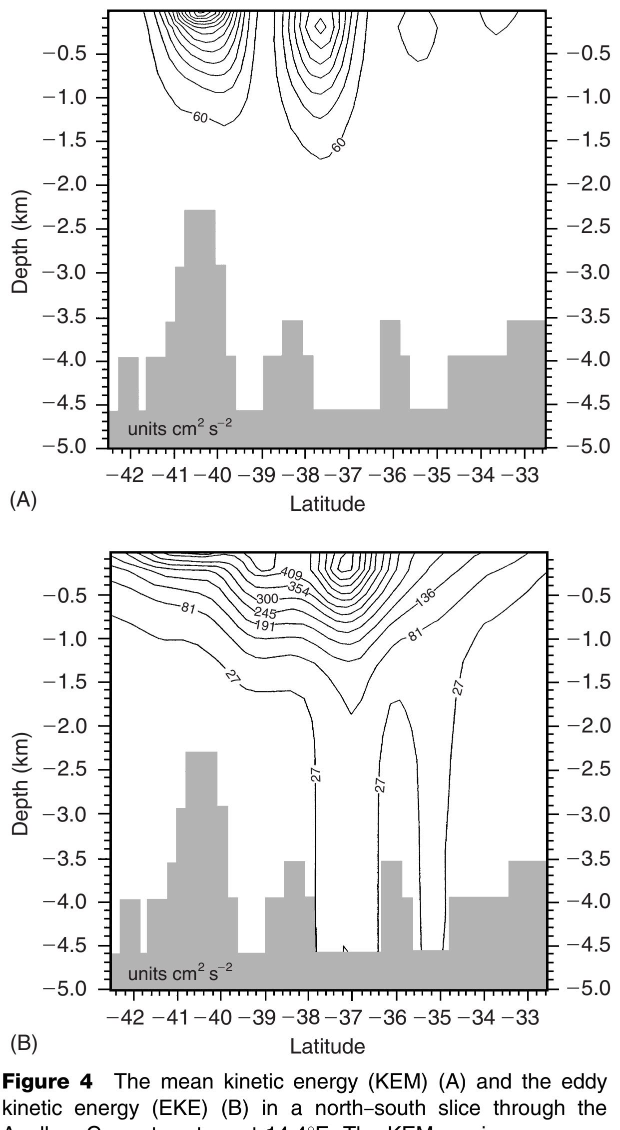

Figure 4 The mean kinetic energy (KEM) (A) and the eddy kinetic energy (EKE) (B) in a north–south slice through the Agulhas Current system at 14.4°E. The KEM maximum corresponds to the mean position of the Agulhas Return Current (Eastward flow) between 40° and 41°S, and the Agulhas Current (Westward flow) between 37°S and 38°S. The EKE distribution is much broader than KEM, which shows the large horizontal extent of the flow variability. The ratio of EKE/KEM is typically about a third, which indicates a very variable current system. (Reproduced from Wells NC, Ivchenko V and Best SE (2000) Instabilities in the Agulhas Retroflection Current system: A comparative model study. *Journal of Geophysical Research* 105: 3233–3246.)

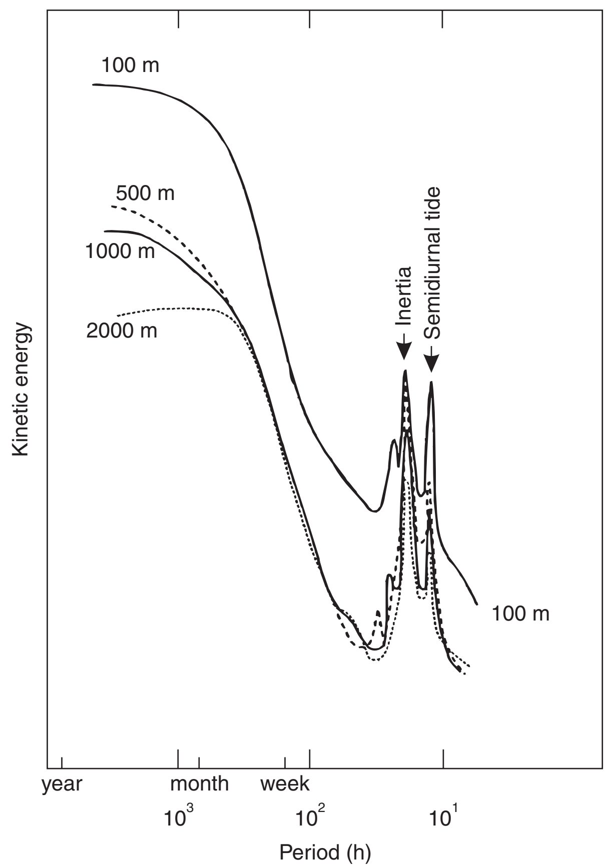

Figure 5 Frequency spectrum of kinetic energy from four depths at site D (39°N, 70°W), north of the Gulf Stream. Note the two high-frequency peaks, coinciding with the inertial period (19 h) and the semidiurnal tide (12.4 h). (Reproduced from Rhines PB (1971) A note on long-period motions at site D. *Deep Sea Research* 18: 21–26.)

inertial frequency is given by $2\Omega \sin \phi$ where Ω is the rotation rate of the earth and ϕ is the latitude. At the lower frequency end, which corresponds to the ocean circulation frequency, there is a broad band of high kinetic energy. This band is due to eddies (*see* **Mesoscale Eddies**) which cause fluctuations of currents on timescales of weeks to months.

For these mean climatological currents, our knowledge has been augmented by the application of the dynamic method. This method is based on the observation that large-scale ocean currents are in geostrophic balance, over large areas of the ocean. Geostrophic balance means that the Coriolis force balances the horizontal pressure gradient force. The geostrophic flow is a good approximation to the flow in the interior of an ocean outside the equatorial region. The horizontal pressure gradient is dependent on the slope of the ocean surface and the horizontal variation of the density distribution within the ocean. In the future, the former may be determined by satellite measurements of the sea surface height and the geoid[1] but at present we do not have an accurate geoid at sufficiently high resolution to measure the sea surface slope. The latter can be determined from temperature, salinity and pressure measurements that have been made over large ocean areas during the last century. The dynamic method allows the determination of the vertical shear of the horizontal geostrophic current, and therefore to determine the absolute geostrophic current, additional measurements are required. For example if the current has been measured at a particular depth then the dynamic method can be referenced to that depth and the vertical profile of current can be obtained.

The recent World Ocean Circulation Experiment (WOCE) hydrographic program has provided more measurements of the ocean than all previous hydrographic programs and will give the most comprehensive assessment of climatological horizontal ocean flow to date.

Recall that the vertical circulation of the ocean cannot be measured directly because it is technically too difficult. Current indirect methods used to determine the vertical circulation rely on the use of mathematical approaches, such as dynamical models, or the use of chemical tracers.

Observations of temperature and salinity can be inserted into a mathematical ocean general circulation model (*see* below and **Elemental Distribution: Overview**) which allows the three-dimensional, circulation to be determined, subject to limitations in the accuracy of the model.

Chemical tracers have been inadvertently injected into the ocean from nuclear tests in the 1960s and from industrial processes (e.g., chlorofluorocarbons) (*see* **CFCs in the Ocean**). Naturally occurring tracers such as ^{14}C also exist. These tracers can be measured with high accuracy in a few laboratories around the world and from their distributions at different times, the three-dimensional circulation can be estimated. This method reveals the time history of the ocean circulation wherever the tracer is measured (*see* **Tritium–Helium Dating**). This is very different information from that provided by the methods previously discussed, but nonetheless it can reveal unique aspects of the flow. For example, nuclear fallout deposited in the surface layer of the Nordic seas in the 1960s was located in the deep western boundary current 10 years later.

An Ocean General Circulation Model

An ocean general circulation model is composed of a set of mathematical equations which describe the time-dependent dynamical flows in an ocean basin. The basin is discretized into a set of boxes of regular horizontal dimensions but variable thickness in the vertical dimension. The horizontal flow (northward and eastward components) is predicted by the momentum equation (**Figure 6A**) at the corners of each box (**Figure 7**).

The forcing for the flow may come from the wind stress (the frictional term in the momentum equation) and from the surface buoyancy fluxes, arising from heat and freshwater (precipitation + runoff–evaporation) exchange with the atmosphere and adjacent landmasses. These buoyancy fluxes change the temperature and salinity in the surface layer of the ocean. The surface water masses are then subducted into the interior of the ocean by the vertical and horizontal components of the flow, where they are mixed with other water masses.

The processes of transport and mixing are described by the temperature and salinity equations (**Figures 6B** and **C**), at the center of each ocean box (**Figure 7**). From these two equations the seawater density, and thence the pressure can be obtained for each box. The horizontal pressure gradient is then determined for the momentum equation, and the vertical velocity is calculated from the horizontal divergence of the flow. This set of time-dependent equations can then be used to describe all the dynamical components of the flow field, provided that suitable initial and boundary conditions are specified.

[1] The geoid is an equipotential surface, which would be represented by the sea level of a stationary ocean. Ocean currents cause deviations in sea level from the geoid.

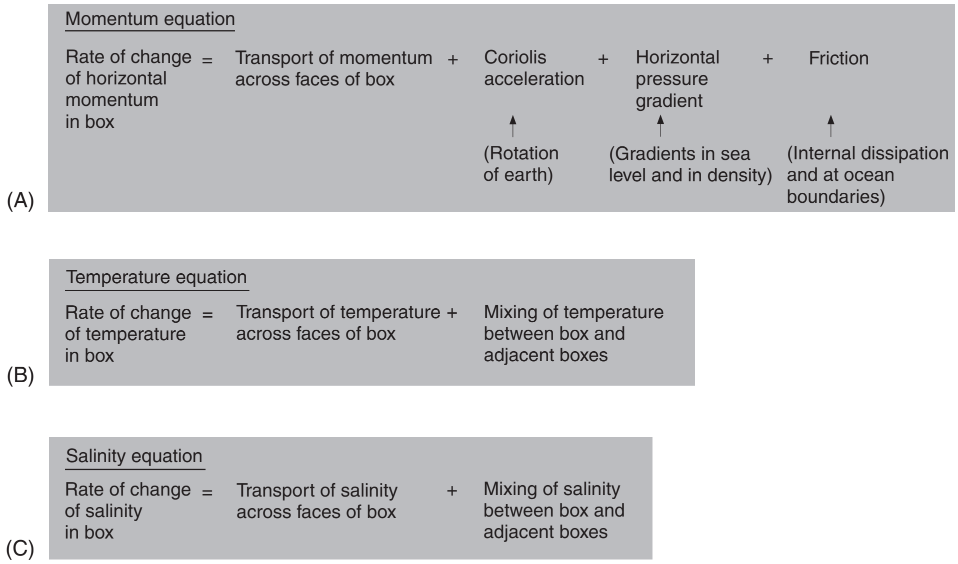

Figure 6 The basic equations for an ocean general circulation model. (A) Momentum equation; (B) temperature equation; (C) salinity equation. (Reproduced from Summerhayes and Thorpe, 1996.)

Wind-driven and Thermohaline Circulation

The wind-driven circulation (*see* **Wind Driven Circulation**) is considered first. The surface layer of the ocean is directly driven by the surface wind stress and is also subject to the exchange of heat and fresh water between ocean and atmosphere. This layer, which is typically less than 100 m in depth, is referred to as the Ekman layer. That is a steady wind stress causes a transport of the surface water 90° to the right of the wind direction in the Northern Hemisphere and 90° to the left in the Southern Hemisphere. This is due to the combined action of the wind stress on the ocean surface and the Coriolis force. These Ekman flows can converge and produce a downwelling flow into the interior of the ocean. Conversely a divergent Ekman transport will produce an upwelling flow from the interior into the surface layer.

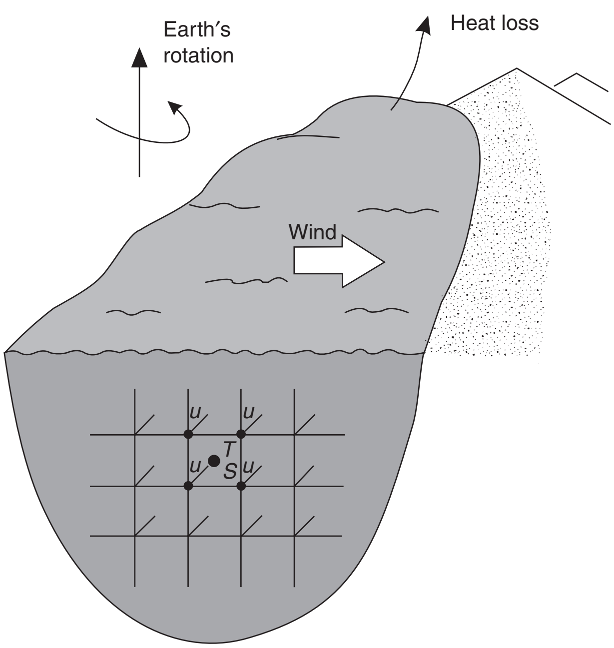

Figure 7 A schematic of the model boxes in an ocean general circulation model. The equations (**Figure 6**) for momentum are solved at the corners of the boxes (u), while the temperature (T), and salinity (S) equations are solved at the centers of boxes. The model is forced by climatological wind stress, surface heat fluxes, and freshwater fluxes. (Reproduced from Summerhayes and Thorpe, 1996.)

This type of flow is known as Ekman pumping, and is directly related to the Curl of the Wind Stress (*see* **Wind Driven Circulation**). It is of fundamental importance for the driving of the large-scale horizontal circulation, in the upper layer of the ocean. For example, between 30° and 50° latitude the climatological westerly wind, drives an Ekman flow equatorward, whereas between 15° and 30° latitude the trade winds drive an Ekman flow polewards. At about 30° latitude the flows converge and sink into the deeper ocean. Before discussing the influence of Ekman pumping on the interior ocean circulation the role of density is considered.

The density of sea water increases with depth. From hydrographic measurements of density, the horizontal variation of the depth of a chosen density surface can be mapped. These constant density surfaces are known as isopycnals. The flow tends to move along these surfaces and therefore the variations in the depth of these surfaces gives a picture of the horizontal flow in the deep ocean, away from the surface layer and benthic layer. The isopycnal surfaces dip down in the center of the subtropical gyre at about 30°. The formation of this lens of light warm water is related to the climatological distribution of surface winds, which produce a convergence of Ekman transport towards the center of the gyre, and a downwelling of surface waters into the interior of the ocean. At the center of the lens, the sea surface domes upwards reaching a height of 1 m above the sea surface at the rim. Due to hydrostatic forces the main thermocline is depressed downwards to depths of the order of 500–1000 m (**Figure 8**).

The surface horizontal circulation flows anticyclonically around the lens with the strongest currents on the western edge, where the slope of the density surface reaches a maximum. These are geostrophic currents, where there is a balance between the Coriolis force and the horizontal pressure gradient force. Generally, the circulation in the subtropical gyres is clockwise in the Northern Hemisphere and anticlockwise in the Southern Hemisphere. These large-scale horizontal gyres are ultimately caused by the climatological surface wind circulation and are found in all the ocean basins.

The surface layer is also subject to heating and cooling, and the exchange of fresh water between ocean and atmosphere, both of which will change the density of the layer. For example, heat is lost over the Gulf Stream on the rim of the light water lens of the subtropical gyre. Recall that flow tends to follow isopycnal layers and these layers will slope downwards towards the center of the gyre. Cooling of the waters in the Gulf Stream leads to the sinking of surface waters to produce a water mass known as 18°C water. This water, which is removed from the surface layer, will slowly move along the isopycnal layers into the thermocline. As it moves clockwise around the gyre it will be subducted in to the deeper layers of the thermocline, in a spiral-like motion (**Figure 9**). The deepest extent of the main thermocline is located in the subtropical gyre to the west of Bermuda on the eastern edge of the Gulf Stream rather than in the center of the ocean basin.

This asymmetry of the gyre is related to the beta effect, i.e., the change of the Coriolis parameter with latitude (*see* **Agulhas Current**).

The subtropical gyres are one of the most well-studied regions of the ocean, and our understanding is therefore most developed in these regions. These gyres occur in the surface and thermocline regions of the ocean and are primarily controlled by the wind circulation, with modifications due to heating and cooling of the surface. The question now arises of why thermoclines are seen in the ocean. For example, why is the warm water not mixed over the whole depth of the ocean and why is the average ocean temperature about 3°C.

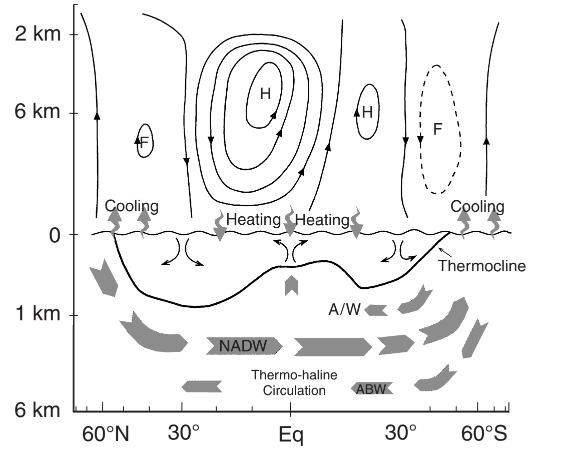

Figure 8 A representation of the meridional average section through the atmosphere for December–February in the Northern Hemisphere. The cells are the Hadley Cell (H) and Ferrel Cell (F). The strength of the cells is represented by the solid contours which are in units of 40 megatonnes/second, whilst the dashed contours are in units of 20 megatonnes/second. Note the predominantly downward motion at ~30 degree latitude, associated with the subtropical anticyclones, and the strong upward motion at equatorial latitudes which is associated with the Inter-Tropical Convergence Zone. A meridional transect through the Atlantic Ocean, showing the position of the main thermocline. The small arrows represent the wind driven downwelling (Ekman pumping) at ~30 degree latitude, and the equatorial upwelling, which occurs within and above the main thermocline. The North Atlantic Deep Water (NADW) is produced in the Labrador and Nordic Seas and is the predominant deep water mass by volume. The Antarctic Intermediate Water (AIW) is produced at ~50°S, and by virtue of its salinity is lighter than the NADW. In contrast Antarctic Bottom Water is the most dense water mass in the worlds ocean and is formed in the Weddell and Ross Seas. These deep flows form part of the thermo-haline circulation. The vertical scales are exaggerated in the lower troposphere and in the upper ocean. The horizontal scale is proportional to the area of the Earth's surface between latitude circles.

To explain the observed behavior thermohaline circulation (*see* **Thermohaline Circulation**), which is generated by small horizontal differences in density, due to temperature and salinity, between low and high latitude is considered. How does it work? Consider an ocean of uniform depth and bounded at the equator and at a polar latitude. We will assume it has initially a uniform temperature and is motionless (for the moment the effect of salinity on density are ignored). This hypothetical ocean is then subject to surface heating at low latitudes and surface cooling at high latitudes. In the lower latitudes the warming will spread downwards by diffusion,

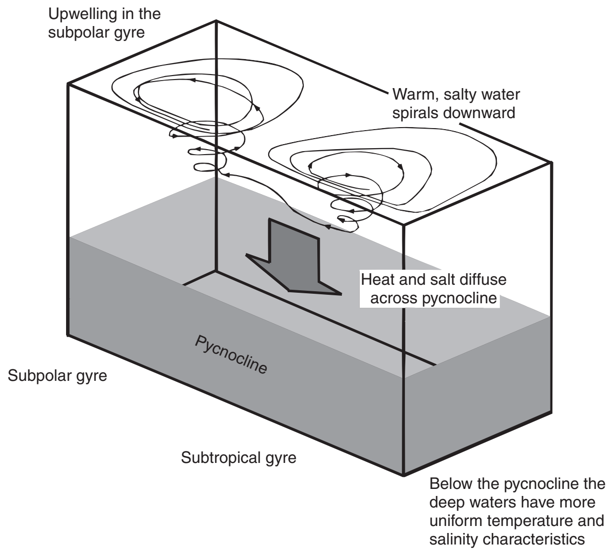

Figure 9 Schematic representation of the wind-driven circulation in the subpolar and subtropical gyre of an ocean basin. The wind circulation causes a convergence of Ekman transport to the center of the subtropical gyre and downwelling into the interior. Conversely in the subpolar gyre there is a divergence of the Ekman transport and upwelling from the interior into the surface layer. This Ekman pumping is responsible for the gyre circulations (see text for details). The western boundary currents are depicted by the closeness of the streamlines. They are caused by the poleward change in the Coriolis force known as the BETA effect. (Reproduced from Bean MS (1997) PhD thesis, University of Southampton.)

whereas in high latitudes the cooling will spread downwards by convection which is a much faster process than diffusion. The heavier colder water will induce a higher hydrostatic pressure at the ocean bottom than will occur at low latitudes. The horizontal pressure gradient at the ocean bottom is directed from the high latitudes to the lower latitudes, and will induce an equatorward abyssal flow of polar water. The flow can not move through the equatorial boundary of our hypothetical ocean and therefore will upwell into the upper layer of the tropical ocean, where it will warm by diffusion. The flow will then return polewards to the high latitudes where it will downwell into deepest layers of the ocean to complete the circuit. It is found that the downwelling occurs in narrow regions of the high latitudes whereas upwelling occurs over a very large area of the tropical ocean. This hypothetical ocean demonstrates the key role of the deep horizontal pressure gradient, caused by horizontal variations in density, for driving the flow.

To explain the observed thermohaline circulation, this hypothetical ocean has to be modified to take into account the Coriolis force, which causes the deep abyssal currents to flow in narrow western boundary currents, the effect of salinity on the density (the haline component of the flow), asymmetries in the buoyancy fluxes between the Northern and Southern Hemispheres and the complex bathemetry of the ocean basins.

There follows a descriptive account of the thermohaline circulation. The deepest water masses (*see* **Water Types and Water Masses**) in the ocean have their origin in the polar seas. These seas experience strong cooling of the surface, particularly in the winter seasons. In the North Atlantic, there are connections through the Nordic seas to the Arctic Ocean, from which sea ice flows. Heat energy melts the ice in the North Atlantic and the melt water gives rise to further cooling. There are two effects on the density of the water: cooling increases the density whereas surface freshening, due to ice melt, decreases the density of the water. The former process usually dominates and hence denser waters are produced. These dense cold waters flow into the Atlantic through the East Greenland and West Greenland Currents and then into the Labrador Current. These cold waters mix and sink beneath the warm North Atlantic Current.

In addition to surface polar currents there are also deep ocean currents (*see* **Abyssal Currents**). The cold saline water entering from the Nordic seas mixes as it sinks to the abyssal layers of the ocean and moves southward as a deep current along the western boundary of the Atlantic. This water mass is known as NADW (North Atlantic Deep Water) and it is the most prominent and voluminous of all the deep water masses in the global ocean. It flows into the Antarctic Circumpolar Current from where it flows into the Indian and Pacific Ocean. In addition to NADW, colder denser water, Antarctic Bottom Water (AABW) enters the Southern Ocean from the Antarctic shelf seas. It is not as voluminous as NADW but it flows northwards in the deepest layers into the Atlantic, where it can be distinguished as far north as 30°N. These deep flows upwell into the thermocline and surface waters where they return to the North Atlantic. This global thermohaline circulation has been termed the global conveyor circulation to signify its role in transporting heat and fresh water (**Figures 10** and **11**) around the planet.

How does this circulation explain the thermocline? The rate at which these cold deep abyssal waters are produced can be estimated and it is known for a steady state in the ocean that production has to be balanced by removal. A large-scale upwelling of the abyssal waters into the thermocline produces this removal. Our simple conceptual picture is of warm thermocline water mixing downwards, balanced by a steady upwelling of the cold abyssal layers. Without the upwelling, the warm waters would mix into the deepest layers of the ocean (*see* **Dispersion and Diffusion in the Deep Ocean**).

The Role of Fresh Water in Ocean Circulation

The present discussion has shown that the wind-driven circulation and the thermohaline circulation are major components of ocean circulation, which are ultimately driven by the surface wind stress and buoyancy fluxes. Buoyancy fluxes are the net effect of heat exchange and the freshwater exchange with the overlying atmosphere. It has been shown that heat exchange is a major process explaining existence of both the thermocline and the deep abyssal water but what is the role of the fresh water in ocean circulation?

In the subtropics there is net removal of fresh water by evaporation. This increases the salinity of the water which, in turn, increases the density of the thermocline waters. Normally this effect is opposed by heating, which lightens the water. However, in the Mediterranean and the Red Sea evaporation produces salient waters, which by virtue of their salinity and cooling in winter, sink to the deepest layers of the basins. At the Straits of Gibraltar (*see* **Flows in Straits and Channels**), this dense saline layer flows out beneath the incoming fresher and cooler Atlantic water. This Mediterranean water forms a distinct layer of high salinity water in the eastern Atlantic Ocean. Similar behavior occurs at Bab el Mandeb adjacent to the Gulf of Aden.

The influence of fresh water is more substantial in the polar oceans. A given amount of fresh water will have a greater effect on density at low temperatures than at high temperatures, because the thermal expansion of sea water decreases with decreasing temperature. At higher latitudes there is a net addition of fresh water into the oceans, which arises from the excess of precipitation over evaporation and the melting of sea ice moving towards the equator from the polar regions.

The addition of fresh water adds buoyancy to the surface layer while cooling removes buoyancy, therefore the fresh water will tend to reduce the effect of the cooling. In the Arctic Ocean (*see* **Arctic Basin Circulation**) the surface layer is colder but less dense than the warmer layer at ~ 100 m and therefore is in equilibrium. This stable halocline in the Arctic Ocean reduces the vertical heat flux in to the deep ocean.

In the subpolar oceans, the addition of fresh water reduces the density of the surface layer and can reduce the prevalence of deep convection. This happened in the late 1960s when fresh water, probably from excessive ice in the Arctic Basin, melted in the subpolar gyre. The effect on the thermohaline circulation is unknown, but it is believed from modeling studies that the decrease in the production of deep waters reduced the thermohaline circulation of the ocean.

What Are the Consequences of This Circulation on the Climate System?

The effects of the ocean circulation on the climate can be understood in terms of the heat capacity of the ocean. The heat capacity of a column of sea water only 2.6 m deep is equivalent to that of a column of whole atmosphere and therefore the ocean heats and cools on a long timescale compared with the atmosphere.

It is known that there is a poleward gradient of temperature, which is driven by the thermal radiation imbalance between the low and high latitudes. In response to this temperature gradient there is a flow of heat from the warmer to cooler latitudes. Both the atmosphere and ocean circulations transfer this heat from low to high latitudes by a variety of mechanisms. In the low latitudes of the atmosphere there is the Hadley cell which transfers low temperature air in the lowest levels via the trade winds towards the equator and transfers warmer air poleward in the upper troposphere (**Figure 8**). At higher latitudes anticyclones and cyclones and their accompanying upper air jet streams transfer heat polewards. In the ocean, the wind-driven Ekman currents transfer heat as surface waters move across latitude circles. This water is returned deeper in the ocean at a different temperature from that of the surface water. The ocean gyres carry heat towards higher latitudes since the poleward flows of the western part of the gyres are warmer than the equatorward flows in the eastern parts of the gyre. Finally, and not least, is the contribution of the thermohaline circulation, which transports warm surface and thermocline waters to the highest latitudes and returns cold water to lower latitudes. **Figures 10** and **11** show the heat transport and fresh water transport in the ocean.

A major difference between the atmosphere and ocean is the relative speed of their circulation. The atmosphere circulation is a fast system, responding on timescales of days to weeks. For example, weather systems in temperate latitudes grow and decay on timescales of a few days. By contrast, the ocean tends to be slower in its response. The fastest

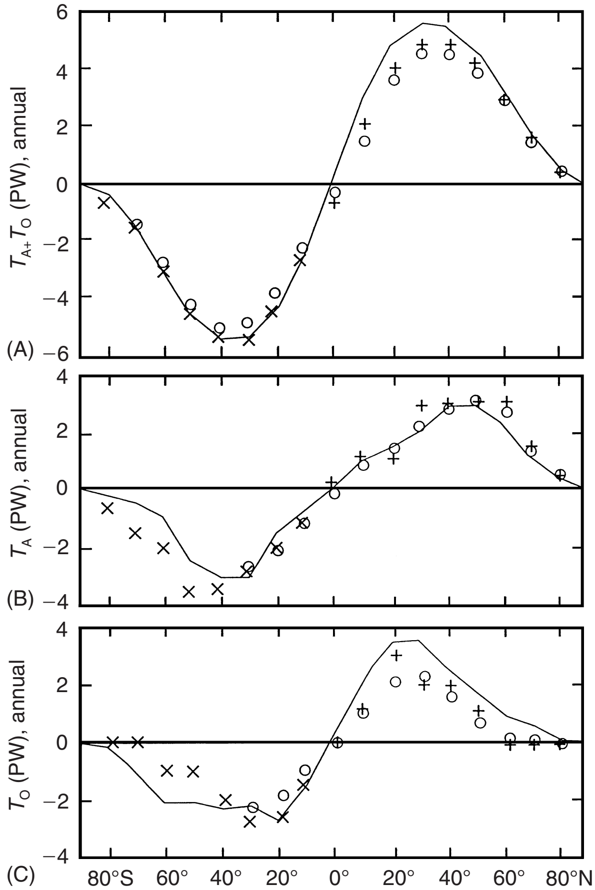

Figure 10 Poleward transfer of heat by (A) ocean and atmosphere together ($T_A + T_O$), (B) atmosphere alone (T_A), and (C) ocean alone (T_O). The total heat transfer (A) is derived from satellite measurements at the top of the atmosphere, that of the atmosphere alone (B) is obtained from measurements of the atmosphere, and (C) is calculated as the difference between (A) and (B). (1 Petawatt (PW) = 10^{15} W.) Data compiled from three sources. (Reproduced from Carrissimo BC, Oort AH and Von der Harr TH (1985) Estimating the meridional energy transports in the atmosphere and ocean. *Journal of Physical Oceanography* 15: 52–91.)

part of the system are the surface Ekman layers which respond to changes in the surface wind circulation on a timescale of one or two days. Changes in wind circulation can cause planetary waves which will change sea level and surface temperature on monthly to seasonal timescales. In particular, the equatorial oceans respond to the surface wind stress on seasonal timescales, which allows a strong coupling between the ocean and atmosphere to take place. This gives rise to phenomena such as the El Niño Southern Oscillation (*see* **El Niño Southern Oscillation (ENSO)**). The subtropical gyres respond to changes in the wind circulation on decadal timescales, whereas the deep thermohaline circulation respond on millenial timescales. There is some evidence for rapid changes of local parts of the thermohaline circulation on timescales 50 years (*see* **Abrupt Climate Change**).

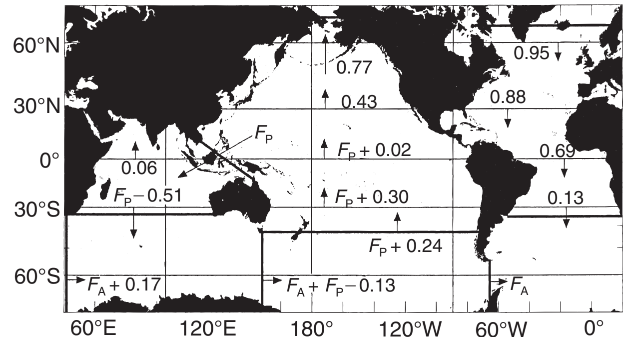

Figure 11 An estimate of the transfer of fresh water ($\times 10^9$ kg s^{-1}) in the world ocean. In polar and equatorial regions precipitation and river run-off exceed evaporation, and hence there is an excess of fresh water, whereas in subtropical regions there is a water deficit. A horizontal transfer of fresh water is therefore required between regions of surplus to regions of deficit. F_P and F_A refer to the freshwater fluxes of the Pacific–Indian throughflow and of the Antarctic Circumpolar Current in the Drake Passage, respectively. (Reproduced from Wijffels SE, Schmitt R, Bryden H and Stigebrandt A (1992) Transport of freshwater by the oceans. *Journal of Physical Oceanography* 22: 155–162.)

Observations of the deep ocean are far fewer in number than at the ocean and land surface. The longest continuous data set is a deep station at Bermuda that commenced operations in 1954. Observations from cruises in the earlier part of the century are of unknown quality and therefore it is difficult to know whether differences are due to the use of different instruments or to real changes in the ocean. It is only since the 1950s that such changes have been accurately measured. **Figure 12** shows changes in the temperature for that period of time across the Atlantic. These changes are of the order of a few tenths of a degree over periods of 15 years. As the heat capacity of the oceans is very much larger than that of the atmosphere, these changes in temperature involve very significant changes in the heat content of the ocean. The World Ocean Circulation Experiment from 1990 to 1997 has provided measurements of ocean properties such as temperature, salinity, and chemical tracers as well as current measurements on a global scale. This set of high quality measurements will provide the baseline from which future changes in ocean circulation can be determined.

Despite the brief record of deep ocean observations, sea surface temperature measurements of reasonable quality go back to the late nineteenth century. These measurements can be used to assess changes in the surface layers (*see* **Heat Transport and Climate**). Salinity measurements are fewer and

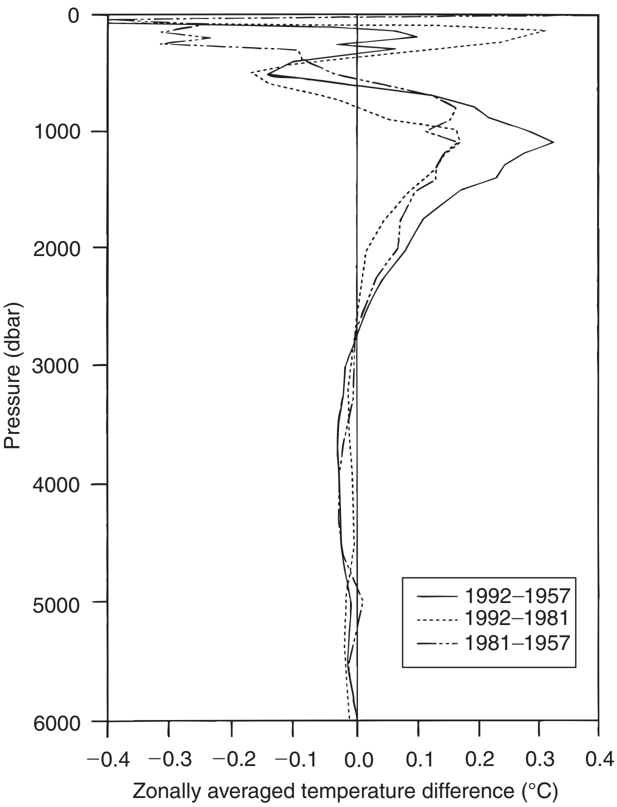

Figure 12 Temperature changes (°C) in the subtropical North Atlantic (24°N), 1957–1992. The measurements have been averaged across 24°N between North Africa and Florida. (Reproduced from Parrilla G, Lavin A, Bryden H, Garcia M and Millard R (1994) Rising temperatures in the subtropical North Atlantic. *Nature* 369: 48–51.)

not as reliable but, nevertheless, changes can be still detected (*see* **Freshwater Transport and Climate**).

Salinity measurements in the Mediterranean over the last century have shown a warming of the Western Mediterranean Deep Water of 0.1°C and increase of 0.05 in salinity. The reasons for this change are not known, but it has been speculated that the change in salinity may be attributed to a reduction in the freshwater flow due to the damming of the Nile and of rivers flowing into the Black Sea.

An important recently identified question is the stability of the thermohaline circulation. The thermohaline circulation is driven by small density differences and therefore changes in the temperature and salinity arising from global warming may alter the thermohaline circulation. In particular, theoretical modeling of the ocean circulation has shown that the thermohaline circulation may be reduced or turned off completely when significant excess fresh water is added to the subpolar ocean. In the event of thermohaline circulation being significantly reduced or stopped, it may take many centuries before it returns to its present value (*see* **Abrupt Climate Change, North Atlantic Oscillation** (**NAO**)).

In view of the current levels of uncertainty, it is necessary to continue to monitor the ocean circulation, as this will provide the key to the understanding of the present circulation and enhance our ability to predict future changes in circulation.

See also

Abrupt Climate Change. Abyssal Currents. Antarctic Circumpolar Current. Arctic Basin Circulation. CFCs in the Ocean. Current Systems in the Atlantic Ocean. Dispersion and Diffusion in the Deep Ocean. Drifters and Floats. Elemental Distribution: Overview. El Niño Southern Oscillation (ENSO). Florida Current, Gulf Stream and Labrador Current. Flows in Straits and Channels. Freshwater Transport and Climate. General Circulation Models. Heat Transport and Climate. Inverse Models. Moorings. North Atlantic Oscillation (NAO). Regional and Shelf Sea Models. Ships. Thermohaline Circulation. Tritium–Helium Dating. Water Types and Water Masses. Waves on Beaches. Wind Driven Circulation.

Further Reading

Gill AE (1982) *Atmosphere–Ocean Dynamics*. London: Academic Press.

Siedler G, Church J and Gould (2001) *Ocean Circulation and Climate*. London: Academic Press.

Summerhayes CP and Thorpe SA (1996) *Oceanography – An Illustrated Guide*. London: Manson Publishing.

Wells NC (1997) *The Atmosphere and Ocean: A Physical Introduction*, 2nd edn. Chichester: John Wiley.

OCEAN COLOR FROM SATELLITES

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Introduction

The term 'ocean color' refers to the spectral composition of the visible light field that emanates from the ocean. The color of the ocean depends on the