Structure. Upper Ocean Mixing Processes. Upper Ocean Responses to Strong Forcing Events. Upper Ocean Time and Space Variability. Upper Ocean Vertical Structure. Wind and Buoyancy-forced Upper Ocean.

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UPPER OCEAN MEAN HORIZONTAL STRUCTURE

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Introduction

The upper ocean is the most variable, most accessible, and dynamically most active part of the marine environment. Its structure is of interest to many science disciplines. Historically, most studies of the upper ocean focused on its impact on shipping, fisheries, and recreation, involving physical and biological oceanographers and marine chemists. Increased recognition of the ocean's role in climate variability and climate change has led to a growing interest in the upper ocean from meteorologists and climatologists.

In the context of this article the upper ocean is defined as the ocean region from the surface to a depth of 1 km and excludes the shelf regions. Although the upper ocean is small in volume when compared to the world ocean as a whole, it is of fundamental importance for life processes in the sea. It determines the framework for marine life through processes that operate on space scales from millimeters to hundreds of kilometers and on timescales from seconds to seasons. On larger space and timescales, its circulation and water mass renewal processes span typically a few thousand kilometers and several decades, which means that the upper ocean plays an important role in decadal variability of the climate system. (In comparison, circulation and water mass renewal timescales in the deeper ocean are of the order of centuries, and the water masses below the upper ocean are elements of climate change rather than climate variability.)

The upper ocean can be subdivided into two regions. The upper region is controlled by air–sea interaction processes on timescales of less than a few months. It contains the oceanic mixed layer, the seasonal thermocline and, where it exists, the barrier layer. The lower region, known as the permanent thermocline, represents the transition from the upper ocean to the deeper oceanic layers. It extends to about 1 km depth in the subtropics, is somewhat shallower near the equator and absent poleward of the Subtropical Front. These elements of the upper ocean will be defined and described in more detail, following an introductory overview of some elementary property fields.

Horizontal Property Fields

The annual mean sea surface temperature (SST) is determined by the heat exchange between ocean and atmosphere. If local solar heat input would be the only determinant, contours of constant SST would extend zonally around the globe, with highest values at the equator and lowest values at the poles. The actual SST field (**Figure 1**) comes close to this simple distribution. Notable departures occur for two reasons.

1. Strong meridional currents transport warm water poleward in the western boundary currents along the east coasts of continents. Examples are the Gulf Stream in the North Atlantic Ocean and the Kuroshio in the North Pacific Ocean. In contrast, cold water is transported equatorward along the west coast of continents.

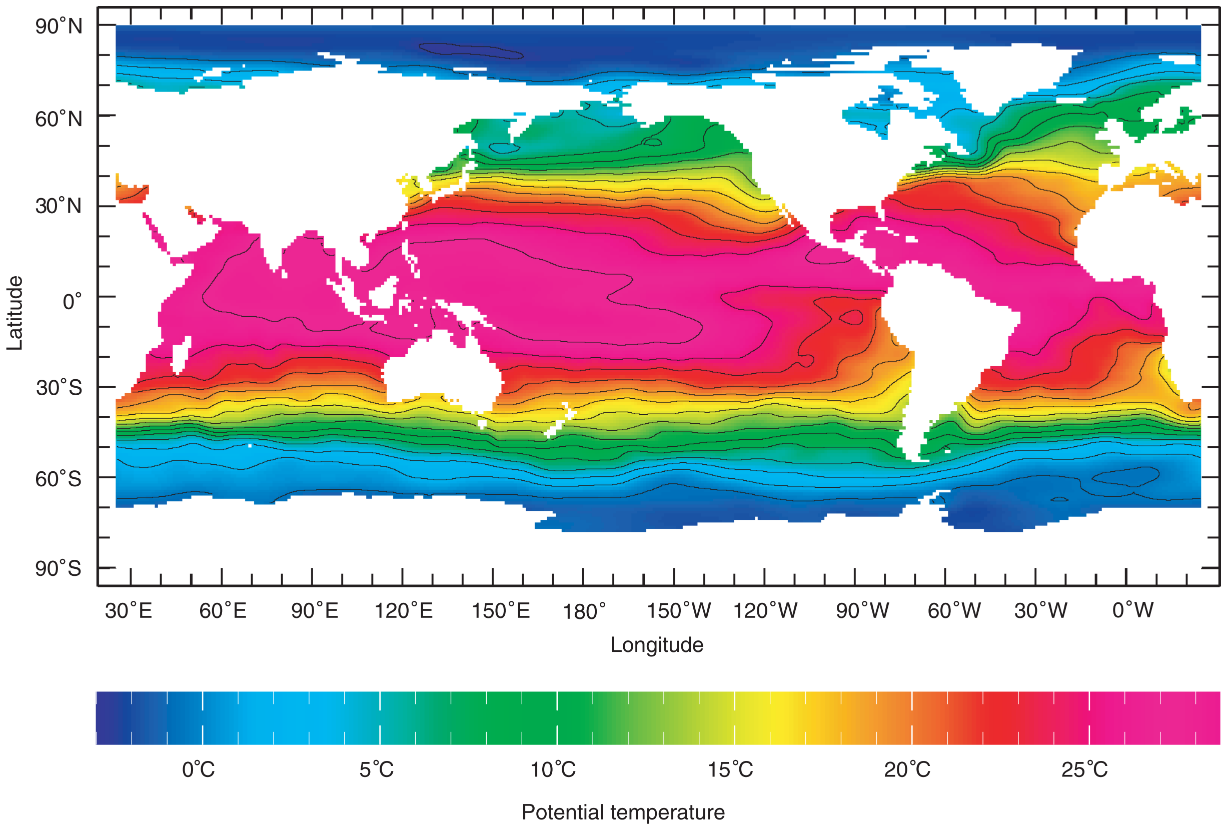


Figure 1 Annual mean sea surface temperature (°C) (the contour interval is 2°C). (Reproduced from *World Ocean Atlas 1994.*)

2. In coastal upwelling regions, for example off the coasts of Peru and Chile or Namibia, SST is lowered as cold water is brought to the surface from several hundreds of meters depth.

The annual mean sea surface salinity (SSS) is controlled by the exchange of fresh water between ocean and atmosphere and reflects it closely (**Figure 2**), the only departures being observed as a result of seasonal ice melting in the polar regions. As a result, the subtropics with their high evaporation and low rainfall are characterized by high salinities, while the regions of the westerly wind systems with their frequent rain-bearing storms are associated with low salinities (**Figure 3**). Persistent rainfall in the intertropical convergence zone produces a regional minimum in the SSS distribution near the equator. Departures from a strict zonal distribution are again observed, for the same reasons listed for the SST distribution. In addition, extreme evaporation rates in the vicinity of large deserts are reflected in high SSS, and large river run-off produced by monsoonal rainfall over south east Asia results in low SSS in the Gulf of Bengal. As a result, the SSS distribution of the north-west Indian Ocean shows a distinct departure from the normal zonal distribution.

Seasonal variations of SST and SSS are mainly due to three factors.

1. Variations in heat and freshwater exchange between ocean and atmosphere are significant for the SST distribution, which shows a drop of SST

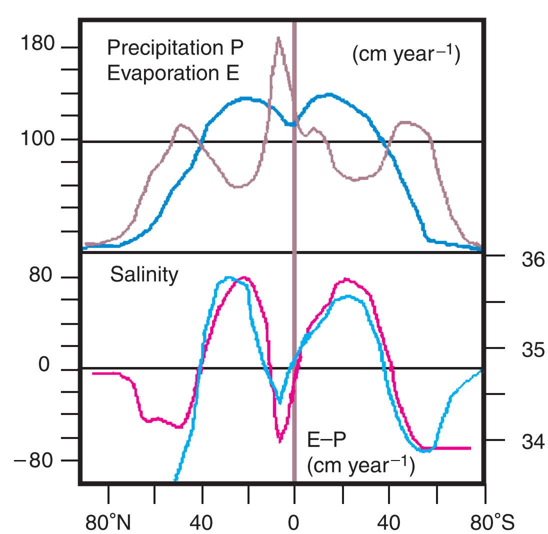


Figure 2 Mean meridional distribution of sea surface salinity and mean meridional freshwater balance (evaporation – precipitation).

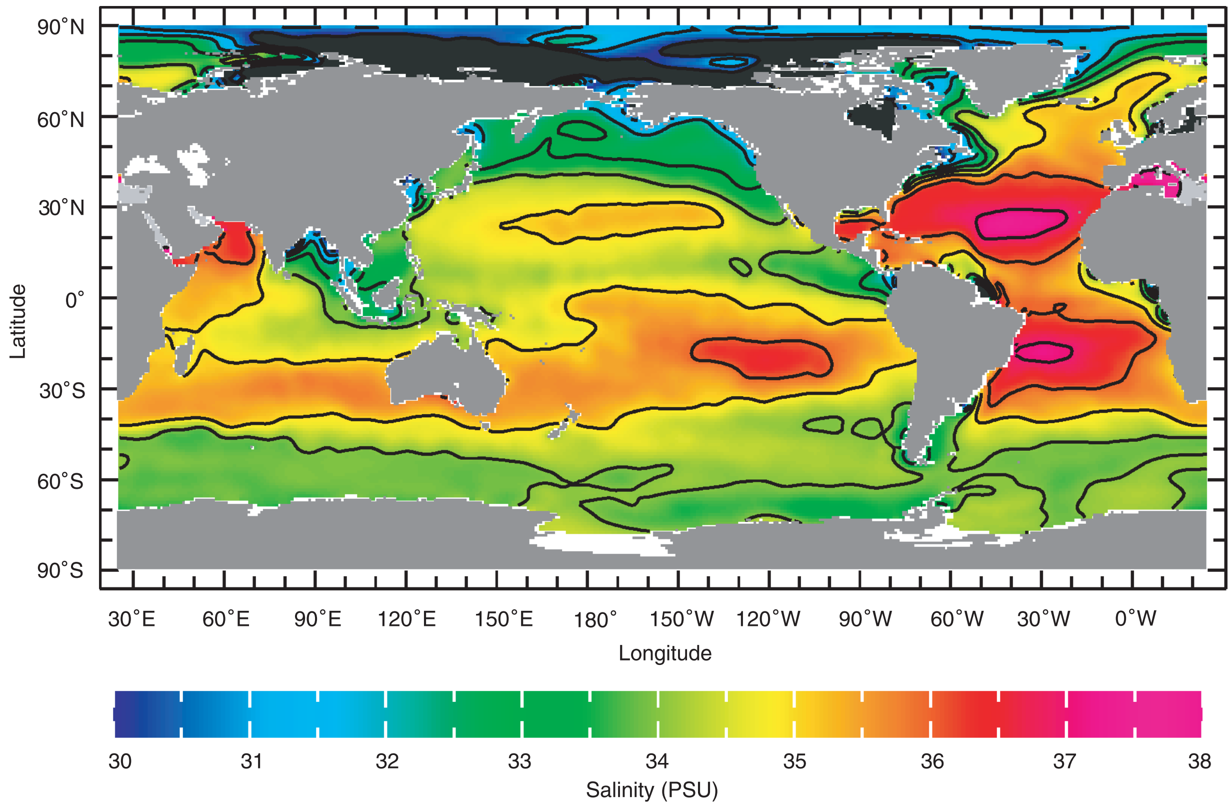


Figure 3 Annual mean sea surface salinity. (Reproduced from *World Ocean Atlas 1994*.)

in winter and a rise in summer, but much less important for the SSS distribution, since rainfall and evaporation do not vary much over the year in most ocean regions.
2. Changes in the ocean current system, particularly in monsoonal regions where currents reverse twice a year, cause the water of some regions to be replaced by water of different SST and SSS.
3. Monsoonal variations of freshwater input from major rivers influences SSS regionally.

The temperature distribution at 500 m depth (**Figure 4**) reflects the circulation of the upper ocean. At this depth the temperature shows little horizontal variation around a mean of 8–10°C. Departures from this mean temperature are, however, observed. (1) The western basins of the subtropics have the highest temperatures in all oceans. They indicate the centres of the subtropical gyres (see below). (2) Poleward of 35° latitude temperatures fall rapidly as the polar regions are reached, an indication of the absence of the permanent thermocline (see below).

The salinity distribution at 500 m depth (**Figure 5**) shows clear similarities to the temperature distribution and a strong correlation between high temperatures and high salinities. The salinity field displays a total range nearly as large as the range seen at the surface (**Figure 3**). The mean salinity varies strongly between ocean basins, with the North Atlantic Ocean having the highest salinity at this depth and the North Pacific Ocean the lowest.

The horizontal oxygen distribution is chosen to represent conditions for marine life. Nutrient levels are inversely related to oxygen, and although the relationship varies between ocean basins, an oxygen maximum can always be interpreted as a nutrient minimum and an oxygen minimum as a nutrient maximum. At the sea surface the ocean is always saturated with oxygen. A map of sea surface oxygen would therefore only illustrate the dependence of the saturation concentration on temperature (and to a minor degree salinity) and show an oxygen concentration of $8\,\mathrm{ml\,l^{-1}}$ or more at temperatures near freezing point and $4\,\mathrm{ml\,l^{-1}}$ at the high temperatures in the equatorial region.

The oxygen distribution at 500 m depth carries a dual signal. It reflects the dependence of the saturation concentration on temperature and salinity in the same way as at the surface but modified by the effect of water mass aging. If water is out of contact with the atmosphere for extended periods of time it experiences an increase in nutrient content from the

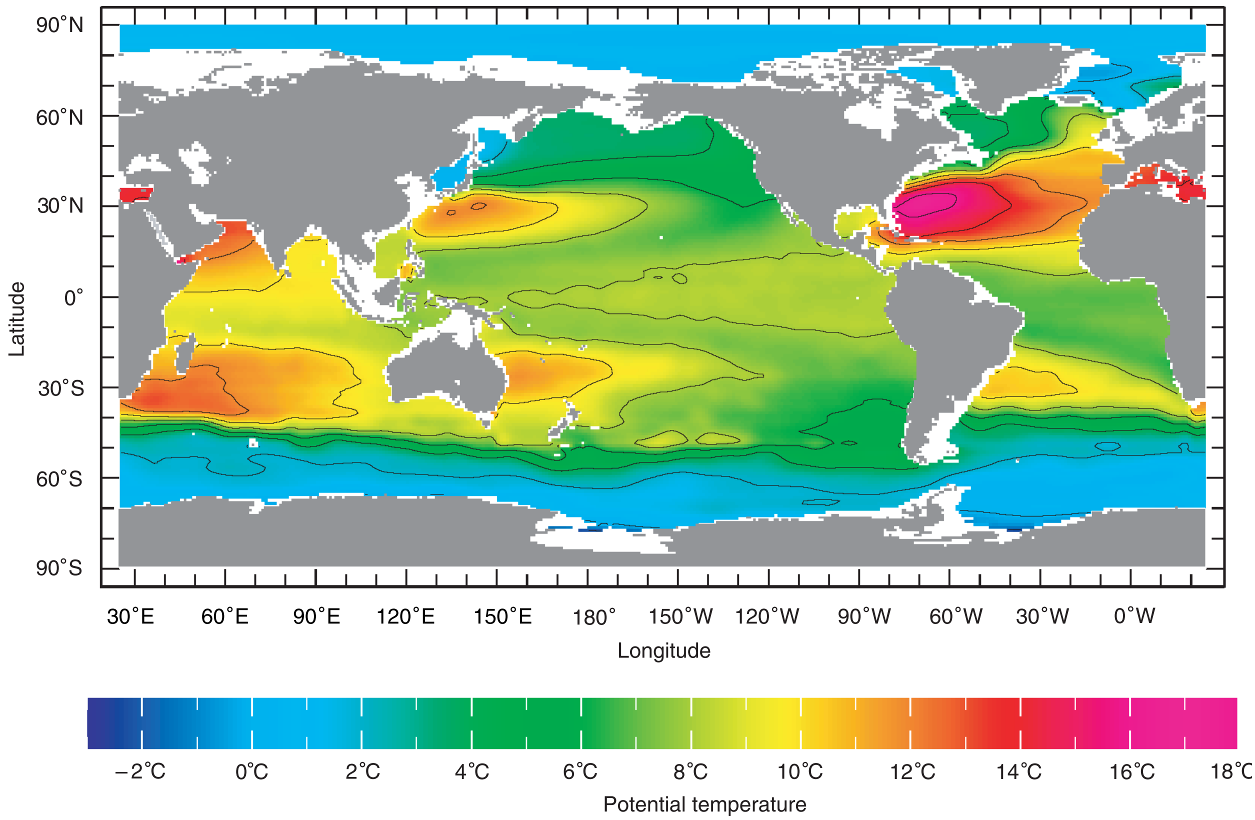


Figure 4 Annual mean potential temperature (°C) at 500 m depth. (Reproduced from *World Ocean Atlas 1994*.)

remineralization of falling detritus; this process consumes oxygen. Water in the permanent thermocline can be a few decades old, which reduces its oxygen content to 60–80% of the saturation value (**Figure 6**). The northern Indian Ocean is an exception to this rule; its long ventilation time (see below) produces oxygen values below 20% saturation. In the polar regions oxygen values at 500 m depth are generally closer to saturation as a result of winter convection in the mixed layer (see below).

The Mixed Layer and Seasonal Thermocline

Exposed to the action of wind and waves, heating and cooling, and evaporation and rainfall, the ocean surface is a region of vigorous mixing. This produces a layer of uniform properties which extends from the surface down as far as the effect of mixing can reach. The vertical extent or thickness of this mixed layer is thus controlled by the time evolution of the mixing processes. It is smallest during spring and summer when the ocean experiences net heat gain (**Figure 7**, **Table 1**). The heat which accumulates at the surface is mixed downward through the action of wind waves. During this period of warming the depth of the mixed layer is determined by the maximum depth which wave mixing can affect. Because winds are often weaker during midsummer than during spring, wind mixing does not reach quite so deep during the summer months, and the mixed layer may consist of two or more layers of uniform properties (**Figure 7**, line 4 of the warming cycle).

During fall and winter the ocean loses heat. This cooling produces a density increase at the sea surface. As a result, mixing during the cooling period is no longer controlled by wave mixing but by convection. The convection depth is determined by the depth to which the layer has to be mixed until static stability is reached. The mixed layer therefore increases with time during fall and winter and reaches its greatest vertical extent just before spring.

The thin region of rapid temperature change below the mixed layer is known as the seasonal thermocline. It is strongest (i.e., is associated with the largest change in temperature) in summer and disappears in winter. In the tropics (within 20° of the equator) the heat loss during winter is not strong enough to erase the seasonal thermocline

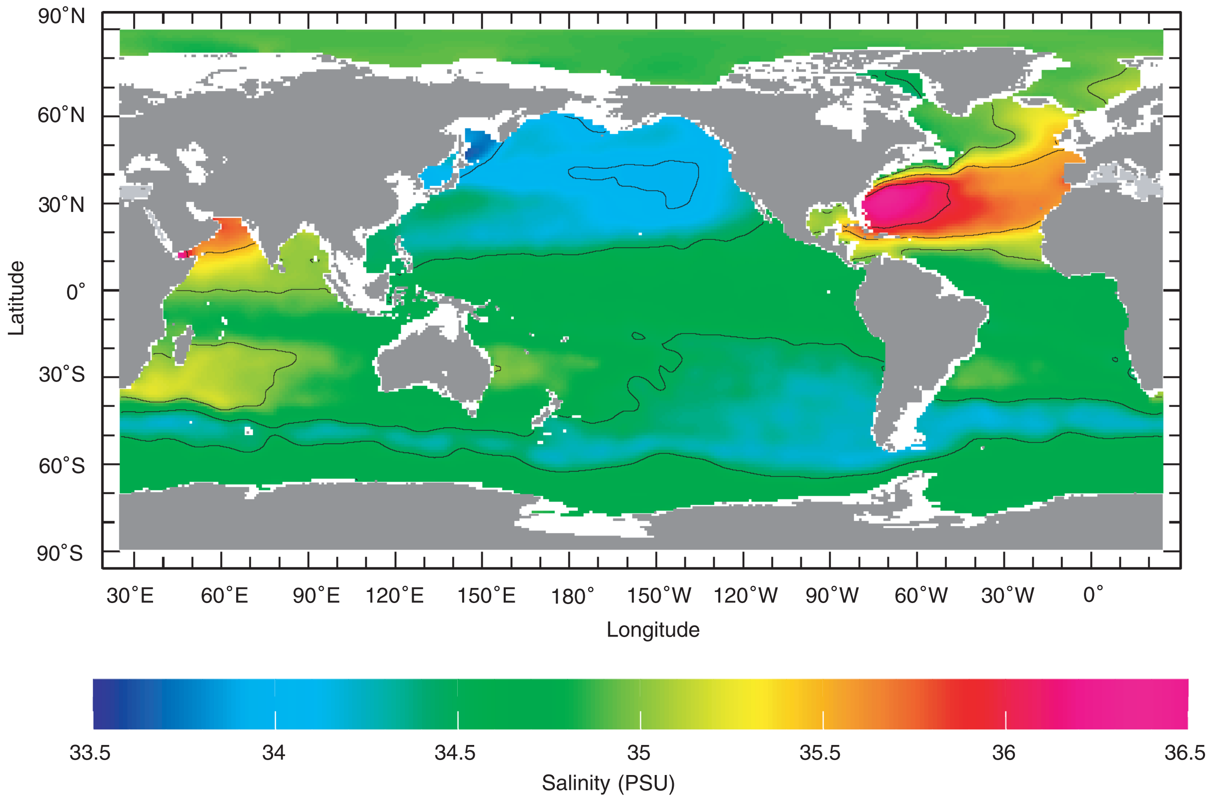


Figure 5 Annual mean salinity (PSU) at 500 m depth. (Reproduced from *World Ocean Atlas 1994*.)

altogether, and the seasonal character of the thermocline is then only seen as a variation of the associated vertical temperature gradient.

In the subtropics the mixed layer depth varies between 20–50 m during summer and 70–120 m during winter. In subpolar regions the mixed layer depth can grow to hundreds of meters during winter. Three locations of particularly deep winter mixed layers are the North Atlantic Ocean between the Bay of Biscay and Iceland, the eastern South Indian Ocean south of the Great Australian Bight and the region to the west of southern Chile. In these regions mixed layer depths can exceed 500 m during late winter.

The Barrier Layer

The mixed layer depth is often equated with the depth of the seasonal thermocline. Historically this view is the result of the paucity of salinity or direct density observations and the resulting need to establish information about the mixed layer from a vertical profile of temperature alone. This approach is acceptable in many situations, particularly in the temperate and subpolar ocean regions. There are, however, situations where it can be quite misleading. A temperature profile obtained in the equatorial western Pacific Ocean, for example, can show uniform temperatures to depths of 80–100 m. Such deep homogeneity in a region where typical wind speeds rarely exceed those of a light breeze cannot be produced by wave mixing.

The truth is revealed in a vertical profile of salinity which shows a distinct salinity change at a much shallower depth, typically 25–50 m, indicating that wave mixing does not penetrate beyond this level and that active mixing is restricted to the upper 25–50 m. In these situations the upper ocean contains an additional layer known as the barrier layer (**Figure 8**). The mixed layer extends to the depth where the first density change is observed. This density change is the result of a salinity increase with depth and therefore associated with a halocline (a layer of rapid vertical salinity change). The temperature above and below the halocline is virtually identical. The barrier layer is the layer between the halocline and the thermocline.

The barrier layer is of immense significance for the oceanic heat budget. In most ocean regions the mixed layer experiences a net heat gain at the sur-

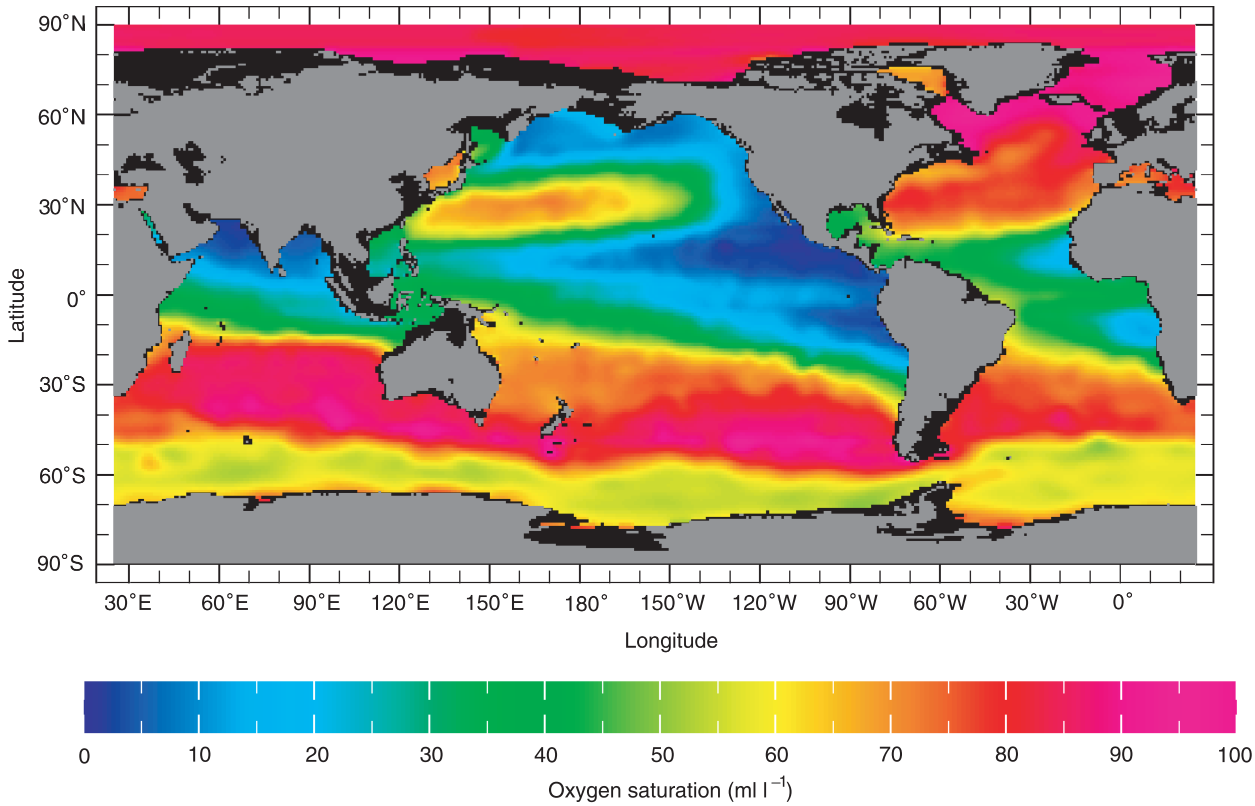


Figure 6 Annual mean oxygen saturation (%) at 500 m depth (the contour interval is 10%). (Reproduced from *World Ocean Atlas* 1994.)

face during spring and summer and has to export heat in order to maintain its temperature in a steady state. If (as described in a previous section) the mixed layer extends down to the seasonal thermocline, this is achieved through the entrainment of colder water into the mixed layer from below. The presence of the barrier layer means that the water

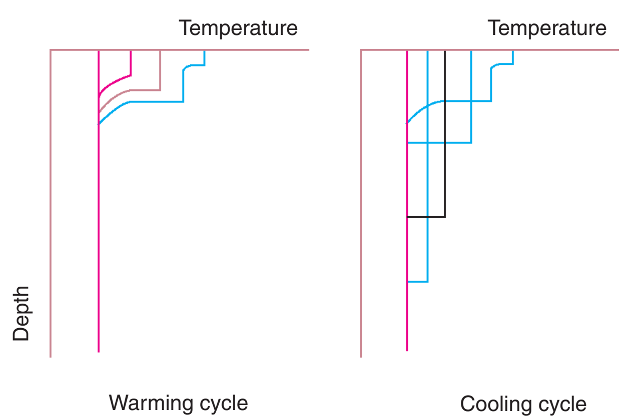


Figure 7 Time evolution of the seasonal mixed layer. Left, the warming cycle; right, the cooling cycle. Numbers can be approximately taken as successive months, with the association shown in **Table 1**.

entrained from the region below the mixed layer is of the same temperature as the water in the mixed layer itself. The entrainment process is still active but does not achieve the necessary heat export. The barrier layer acts as a barrier to the vertical heat flux, and the heat gained by the mixed layer has to be exported through other means, mainly through horizontal advection by ocean currents and, if the mixed layer is sufficiently transparent to the incoming solar radiation, through direct downward heat transfer from the atmosphere to the barrier layer.

The existence of the barrier layer has only come to light in the last decade or two when high-quality salinity measurements became available in greater numbers. It has now been documented for all tropical ocean regions. In the Pacific Ocean the regional extent of the barrier layer is closely linked with high local rainfall in the Intertropical and South Pacific Convergence Zones of the atmosphere. This suggests that the Pacific barrier layer is formed by the lowering of the salinity in the shallow mixed layer in response to local rainfall. In contrast, the barrier layer in the Indian Ocean varies seasonally in extent, and the observed lowering of the mixed layer salinity seems to be related to the spreading of fresh

Table 1 Association between number in **Figure 7** and months

	Northern Hemisphere		*Southern Hemisphere*	
Number in Figure 7	*Warming cycle*	*Cooling cycle*	*Warming cycle*	*Cooling cycle*
1	February	June	August	December
2	April	August	October	February
3	May	September	November	March
4	June	October	December	April
5		January		July

water from rivers during the rainy monsoon season. In the Atlantic Ocean the barrier layer is most likely the result of subduction of high salinity water from the subtropics under the shallow tropical mixed layer. There are also observations of seasonal barrier layers in other tropical ocean regions, such as the South China Sea.

The Subtropical Gyres and the Permanent Thermocline

The permanent or oceanic thermocline is the transition from the upper ocean to the deeper oceanic layers. It is characterized by a relatively rapid decrease of temperature with depth, with a total temperature drop of some 15°C over its vertical extent, which varies from about 800 m in the subtropics to less than 200 m near the equator. This depth range does not display the relatively strong currents experienced in the upper ocean but still forms part of the general wind driven circulation, so its water moves with the same current systems seen at the sea surface but with lesser speed.

The permanent thermocline is connected with the atmosphere through the Subtropical Convergence, broad region of the upper ocean poleward of the subtropics where the wind-driven surface currents converge, forcing water to submerge ('subduct') under the upper ocean layer and enter the permanent thermocline. This convergence is particularly intense in the subtropical front, a region of enhanced horizontal temperature change within the Subtropical Convergence found at about 35°N and 40°S. The Subtropical Front is therefore considered the poleward limit of the permanent thermocline (**Figure 9**).

There is also a zonal variation in the vertical extent, with smallest values in the east and largest values in the west. Taken together, the permanent thermocline appears bowl shaped, being deepest in the western parts of the subtropical ocean (25–30°N

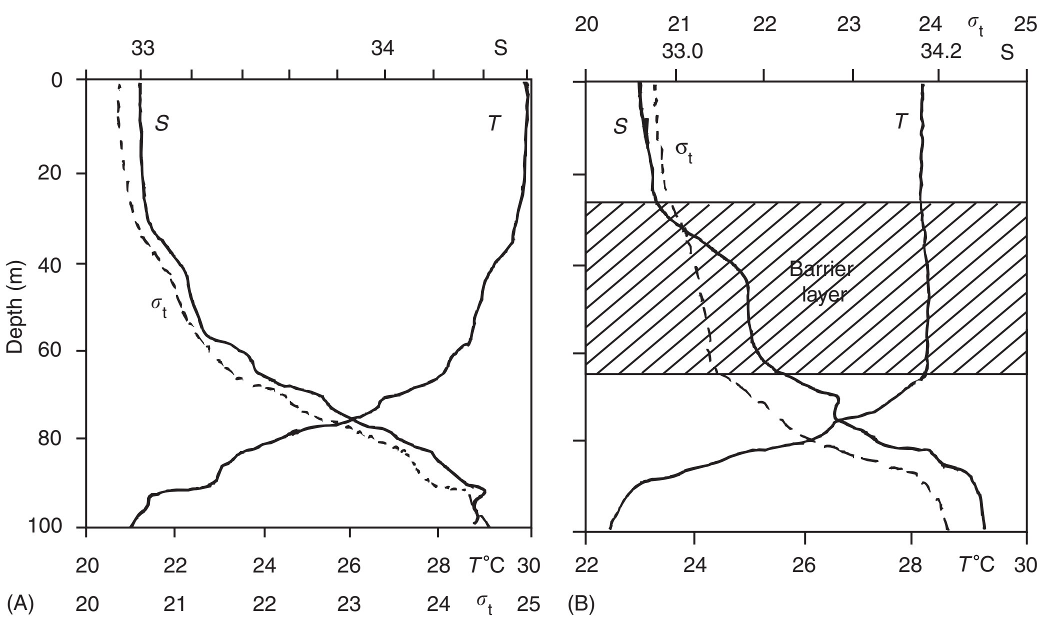


Figure 8 The structure of the upper ocean in the absence (A) and presence (B) of a barrier layer. T: temperature (°C), S: salinity, σ_t: density. Note the uniformity of temperature (*T*) from the surface to the bottom of the barrier layer in (B). The stations were taken in the central South China Sea during September 1994.

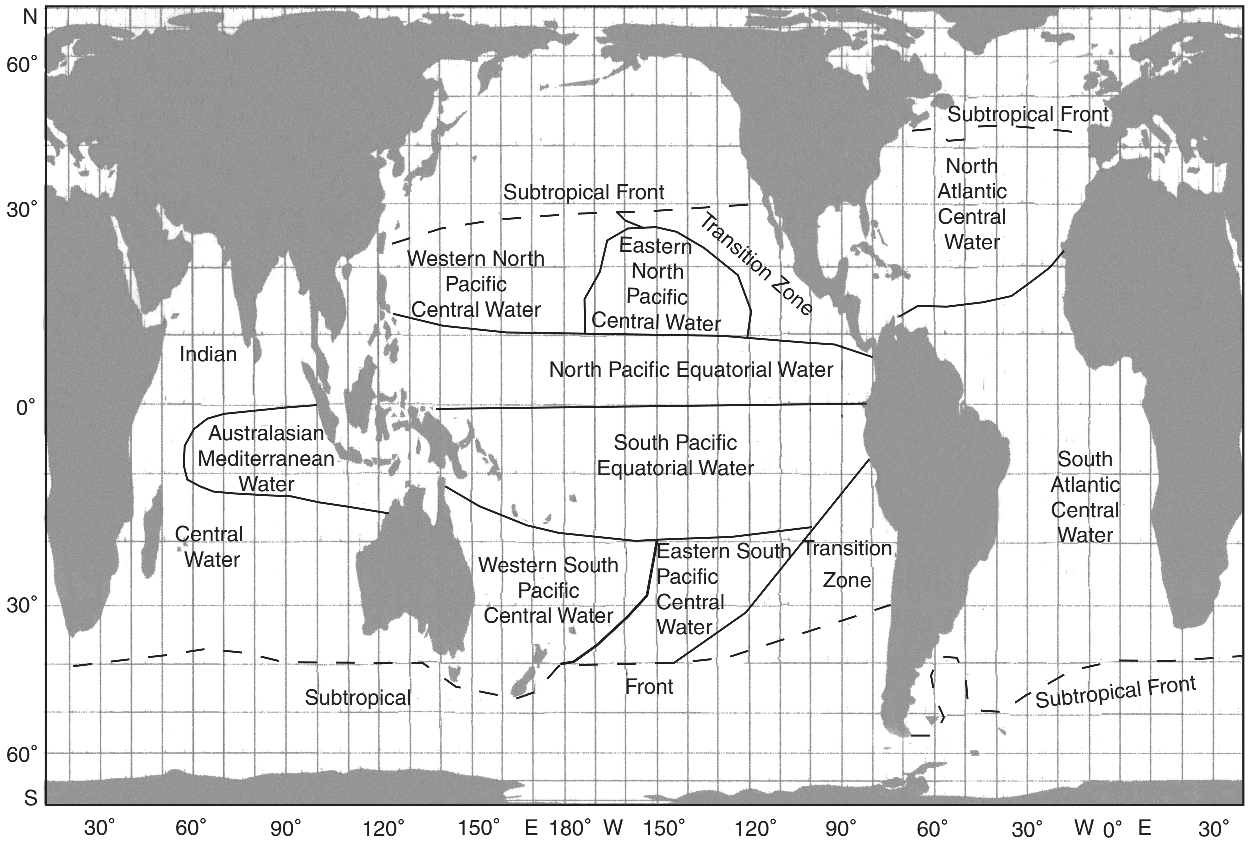


Figure 9 Regional distribution of the water masses of the permanent thermocline.

and 30–35°S). The shape is the result of geostrophic adjustment in the wind-driven circulation, which produces anticyclonic water movement in the subtropics known as the subtropical gyres.

In most ocean regions the permanent thermocline is characterized by a tight temperature–salinity (TS) relationship, lower temperatures being associated with lower salinities. If temperature or salinity is plotted on a constant depth level across the permanent thermocline, the highest temperature and salinity values are found in the western subtropics (**Figures 4** and **5**). The tight TS relationship indicates the presence of a stable water mass, known as Central Water. This water mass is formed at the surface in the subtropical convergence, particularly at the downstream end of the western boundary currents, where it is subducted and from where it renews ('ventilates') the permanent thermocline by circulating in the subtropical gyres, moving equatorward in the east, westward with the equatorial current system and returning to the ventilation region in the west. As a result the age of the Central Water does not increase in a simple meridional direction from the subtropical front towards the equator but is lower in the east and higher in the west.

As the Subtropical Front is a feature of both hemispheres, each ocean, with the exception of the Indian Ocean which does not reach far enough north to have a Subtropical Front in the northern hemisphere, has Central Water of northern and southern origins (**Figure 9**). Fronts between the different varieties of Central Water are a prominent feature of the permanent thermocline. These fronts are characterized by strong horizontal temperature and salinity gradients but relatively small density change because the effect of temperature on density is partly compensated by the effect of salinity. As a result small-scale mixing processes such as double diffusion, filamentation and interleaving are of particular importance in these fronts.

The Equatorial Region

The equatorial current system occupies the region 15°S–15°N and is thus more than 3000 km wide. Most of it is taken up by the North and South Equatorial Currents, the westward flowing equatorial elements of the subtropical gyres discussed above. Between these two currents flows the North Equatorial Countercurrent as a relatively narrow band eastward along 5°N in the Atlantic and Pacific Oceans and, during the north-east monsoon season, along 5°S in the Indian Ocean. Another eastward current, the Equatorial Undercurrent, flows submerged along the equator, where it occupies the depth range 50–250 m as a narrow band of only 200 km width.

Currents near the equator are generally strong, and for dynamical reasons transport across the equator is more or less restricted to the upper mixed layer and to a narrow regime of a few hundred kilometers width along the western boundary of the oceans. This restriction and the narrow eastward currents embedded in the general westward flow, shape the distribution of properties in the permanent thermocline near the equator. In situations where subtropical gyres exist (the Atlantic and Pacific Oceans) in both hemispheres they enter the equatorial current system from the north east and from the south east, leaving a more or less stagnant region ('shadow zone') between them near the eastern coast. **Figure 10** shows the age distribution for the Atlantic Ocean. The presence of particularly old water in the east indicates a stagnant region or 'shadow zone' between the subtropical gyres.

The strong eastward flowing currents in the equatorial current system modify the age distribution in the permanent thermocline further. In **Figure 10** the Equatorial Undercurrent manifests itself as a band of relatively young water, which is carried eastward.

The Indian Ocean does not extend far enough to the north to have a subtropical convergence in the Northern Hemisphere. In the absence of a significant source of thermocline water masses north of the equator the water of the Northern Hemisphere can only be ventilated from the south. **Figure 11**

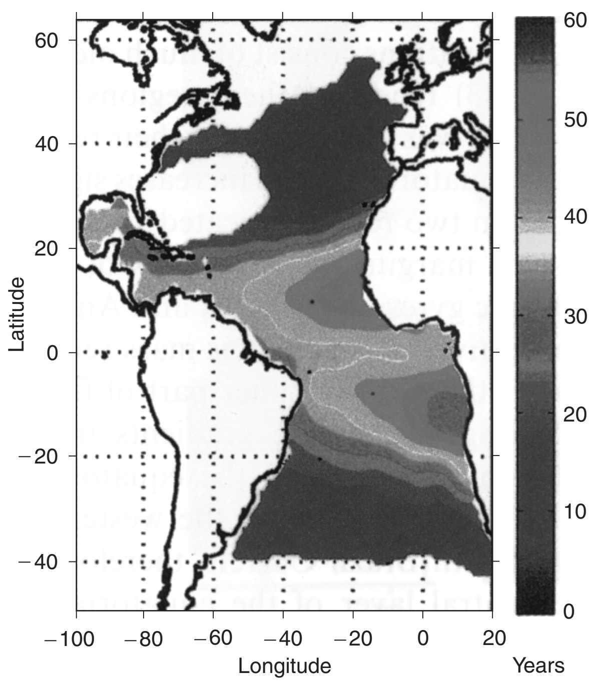


Figure 10 Pseudo age of Central Water in equatorial region of the Atlantic Ocean at 500 m depth. The quantity pseudo age expresses the time elapsed since the water had last contact with the atmosphere; it is determined by using an arbitrary but realistic oxygen consumption rate for the permanent thermocline. (Reproduced from Poole and Tomczak M (19) Optimum multiparameter analysis of the water mass structure in the Atlantic Ocean thermocline. *Deep-Sea Research* 46: 1895–1921.)

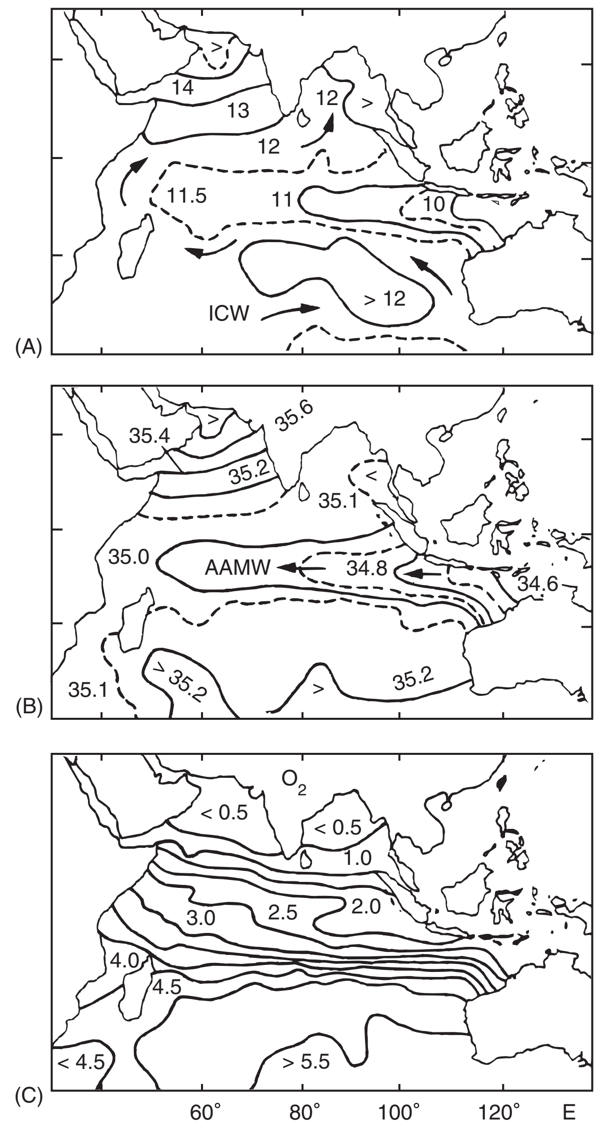


Figure 11 Climatological mean temperature (°C) (A), salinity (PSU) (B) and oxygen concentration (ml l^{-1}) (C) in the Indian Ocean for the depth range 300–450 m, with pathways for Indian Central Water (ICW) and Australasian Mediterranean Water (AAMW). (Reproduced from Tomczak and Godfrey, 1994.)

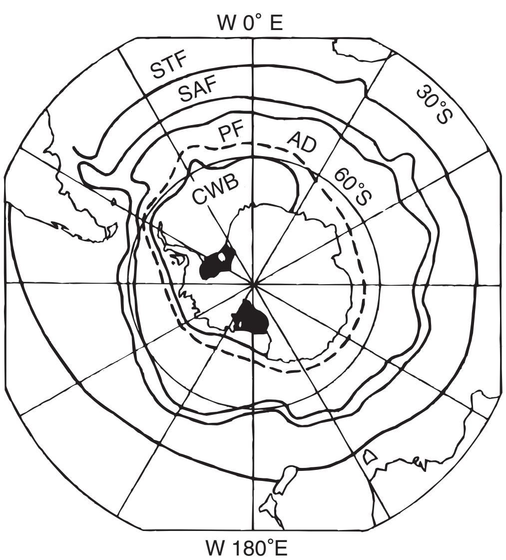
shows property fields of the permanent thermocline in the Indian Ocean and pathways of its water masses. The region between 5°S and the equator is dominated by the westward flow of Australasian Mediterranean Water (AAMW), a water mass formed in the Indonesian seas. Its mass transport is relatively modest, and it is mixed into the surrounding waters before it reaches Africa. Indian Central Water (ICW) originates near 30°S in large volume; it joins the anticyclonic circulation of the subtropical gyre and can be followed (at the depth level shown in **Figure 11** by its temperature of 11.7°C and salinity of 35.1) across the equator along the African coast and into the Northern Hemisphere. The flow into the Northern Hemisphere is thus severely restricted, and the ventilation of the northern Indian Ocean thermocline is unusually inefficient. This is reflected in the extremely low oxygen content throughout the northern Indian Ocean.

The Polar Regions

Poleward of the subtropical front the upper ocean changes character. As polar latitudes are approached the distinction between upper ocean and deeper layers disappears more and more. There is no permanent thermocline; temperature, salinity and all other properties are nearly uniform with depth. The surface mixed layer is, of course, still well defined as the layer affected by wave mixing, but its significance for the heat exchange with the atmosphere is greatly reduced because frequent convection events produced by surface cooling penetrate easily into the waters below the mixed layer.

Because in the polar regions the upper ocean and the deeper layers form a single dynamic unit, the horizontal structure of the upper ocean in these regions is strongly influenced by features of the deeper layers. **Figure 12** shows the arrangement of the various fronts in the Southern Ocean. The fronts are associated with the Antarctic Circumpolar Current. They occupy about 20% of its area but carry 75% of its transport. These fronts extend from the surface to the ocean floor and are thus not exclusive features of the upper ocean.

At the low temperatures experienced in the polar seas the density is very insensitive to temperature changes and is controlled primarily by the salinity. During ice formation salt seeps out and accumulates under the ice, increasing the water density and causing it to sink. Salt from the upper ocean is thus transferred to the deep ocean basins. As a result, a significant amount of fresh water is added to the upper ocean when the ice melts and floats over the oceanic water. The resulting density gradient guarantees stability of the water column even in the presence of temperature inversions. A characteristic feature of the upper ocean in the polar regions is therefore the widespread existence of shallow temperature maxima. In the Arctic Ocean the water below the upper ocean can be as much as 4°C



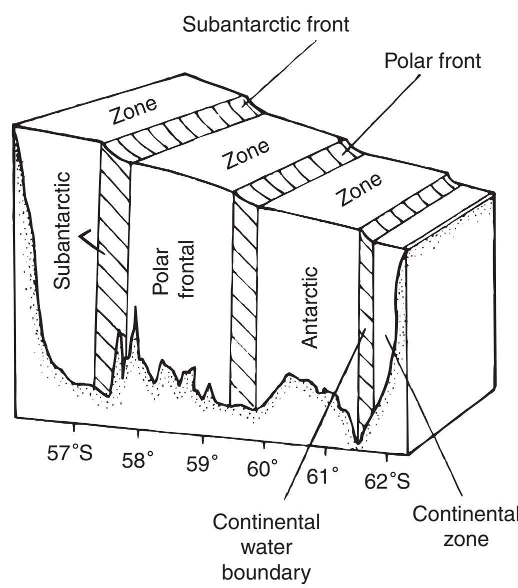


Figure 12 Fronts in the Southern Ocean. (Reproduced from Tomczak and Godfrey, 1994.) STF, Subtropical Front; SAF, Subantarctic Front; PF, Polar Front; CWB, Continental Water boundary; AD, Antarctic Divergence.

warmer than the mixed layer. Intermediate temperature maxima in the Antarctic Ocean are less pronounced (up to 0.5°C) but occur persistently around Antarctica.

See also

Ekman Transport and Pumping. Geophysical Heat Flow. Heat Transport and Climate. Satellite Measurements of Salinity. Satellite Remote Sensing of Sea Surface Temperatures. Thermohaline Circulation. Wind and Buoyancy-forced Upper Ocean. Wind Driven Circulation.

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UPPER OCEAN MIXING PROCESSES

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Introduction

The ocean's effect on weather and climate is governed largely by processes occurring in the few tens of meters of water bordering the ocean surface. For example, water warmed at the surface on a sunny afternoon may remain available to warm the atmosphere that evening, or it may be mixed deeper into the ocean not to emerge for many years, depending on near-surface mixing processes. Local mixing of the upper ocean is predominantly forced from the state of the atmosphere directly above it. The daily cycle of heating and cooling, wind, rain, and changes in temperature and humidity associated with mesoscale weather features produce a hierarchy of physical processes that act and interact to stir the upper ocean. Some of these are well understood, whereas others have defied both observational description and theoretical understanding.

This article begins with an example of *in situ* measurements of upper ocean properties. These observations illustrate the tremendous complexity of the physics, and at the same time reveal some intriguing regularities. We then describe a set of idealized model processes that appear relevant to the observations and in which the underlying physics is understood, at least at a rudimentary level. These idealized processes are first summarized, then discussed individually in greater detail. The article closes with a brief survey of methods for representing upper ocean mixing processes in large-scale ocean models.

Over the past 20 years it has become possible to make intensive turbulence profiling observations that reveal the structure and evolution of upper ocean mixing. An example is shown in **Figure 1**, which illustrates mixed-layer[1] evolution, temperature structure and small-scale turbulence. The small white dots in **Figure 1** indicate the depth above which stratification is neutral or unstable and mixing is intense, and below which stratification is stable and mixing is suppressed. This represents a means of determining the vertical extent of the mixed layer directly forced by local atmospheric conditions. (We will call the mixed-layer depth D.) Following the change in sign from negative (surface heating) to positive (surface cooling) of the surface buoyancy flux, J_b^0, the mixed layer deepens. (J_b^0 represents the flux of density (mass per unit volume) across the sea surface due to the combination of heating/cooling and evaporation/precipitation.) The mixed layer shown in **Figure 1** deepens each night, but the rate of deepening and final depth vary. Each day, following the onset of daytime heating, the mixed layer becomes shallower.

Significant vertical structure is evident within the nocturnal mixed layer. The maximum potential temperature (θ) is found at mid-depth. Above this, θ is smaller and decreases toward the surface at the rate of about 2 mK in 10 m. The adiabatic change in temperature, that due to compression of fluid parcels with increasing depth, is 1 mK in 10 m. The region above the temperature maximum is superadiabatic, and hence prone to convective instability.

[1] Strictly, a mixed layer refers to a layer of fluid which is not stratified (vertical gradients of potential temperature, salinity and potential density, averaged horizontally or in time, are zero). The terminology is most precise in the case of a convectively forced boundary layer. Elsewhere, oceanographers use the term loosely to describe the region of the ocean that responds most directly to surface processes. Late in the day, following periods of strong heating, the mixed layer may be quite shallow (a few meters or less), extending to the diurnal thermocline. In winter and following series of storms, the mixed layer may extend vertically to hundreds of meters, marking the depth of the seasonal thermocline at midlatitudes.