Further Reading

Davis RE, de Szoeke R, Halpern D and Niiler P (1981) Variability in the upper ocean during MILE. Part I: The heat and momentum balances. *Deep-Sea Research* 28: 1427–1452.

Ekman VW (1905) On the influence of the earth's rotation on ocean currents. *Arkiv Matematik, Astronomi och Fysik* 2: 1–52.

Eriksen CC, Weller RA, Rudnick DL, Pollard RT and Regier LA (1991) Ocean frontal variability in the Frontal Air–Sea Interaction Experiment. *Journal of Geophysical Research* 96: 8569–8591.

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

Gregg MC (1989) Scaling turbulent dissipation in the thermocline. *Journal of Geophysical Research* 94: 9686–9698.

Langmuir I (1938) Surface motion of water induced by wind. *Science* 87: 119–123.

Lighthill MJ and Pearce RP (eds) (1981) *Monsoon Dynamics*. Cambridge: Cambridge University Press.

Munk W (1981) Internal waves and small-scale processes. In: Warren BA, Wunsch C (eds) *Evolution of Physical Oceanography*, pp. 264–291. Cambridge, USA: MIT Press.

Philander SG (1990) *El Niño, La Niña, and the Southern Oscillation*. San Diego: Academic Press.

Roden GI (1984) Mesoscale oceanic fronts of the North Pacific. *Annals of Geophysics* 2: 399–410.

UPPER OCEAN VERTICAL STRUCTURE

J. Sprintall, University of California San Diego, La Jolla, CA, USA
M. F. Cronin, NOAA Pacific Marine Environmental Laboratory, Seattle, WA, USA

doi:10.1006/rwos.2001.0149

Introduction

The upper ocean connects the surface forcing from winds, heat, and fresh water, with the quiescent deeper ocean where this heat and fresh water are sequestered and released on longer time and global scales. Classically the surface layer includes both an upper mixed layer, that is subject to the direct influence of the atmosphere, and also the highly stratified zone below the mixed layer, where vertical property gradients are maximum. Although all water within the surface layer has been exposed to the atmosphere at some point in time, water most directly exposed lies within the mixed layer. Thus, the surface layer vertical structure reflects not only immediate changes in the forcing, but also anomalies associated with earlier forcing events. Further, these forcing events may have occurred either locally in the region, or remotely at other locations and transferred by ocean currents. This article first defines the major features of the upper ocean vertical structure and discusses what causes and maintains them. Then the rich variability in the vertical shapes and forms that these structures can assume through variation in the atmospheric forcing is demonstrated.

Major Features of the Upper Ocean Vertical Structure

The vertical structure of the upper ocean is primarily defined by changes in the temperature and salinity, which together control the water column's density structure. Within the upper ocean surface layer, a number of distinct layers can be distinguished that are formed by different processes over different timescales: the upper mixed layer, the seasonal pycnocline, and the permanent pycnocline (**Figure 1**). Right at the ocean surface in the top few millimeters, a cool 'skin' exists with lowered temperature caused by the combined heat losses from long-wave radiation, sensible and latent heat fluxes. The cool skin is always present and is the actual sea surface temperature (SST) measured by airborne infrared radiometers. In contrast, *in-situ* sensors in the top few meters measure the 'bulk' SST. The cool skin temperature is of the order 0.1–0.5 K cooler than the bulk temperature. In terms of the upper ocean buoyancy, the bulk SST is often a more appropriate characterization of the surface temperature. Air–sea fluxes are transported through the molecular layer almost instantaneously so that the upper mixed layer can be considered in direct contact with the atmosphere. For this reason, when defining the depth of the upper ocean structure, the changes in water properties are generally made relative to the bulk SST measurement.

The upper mixed layer is the site of active air–sea exchanges. Energy for the upper mixed layer to change its vertical structure comes from wind mixing or through a surface buoyancy flux. Wind mixing causes the vertical turbulence and convective

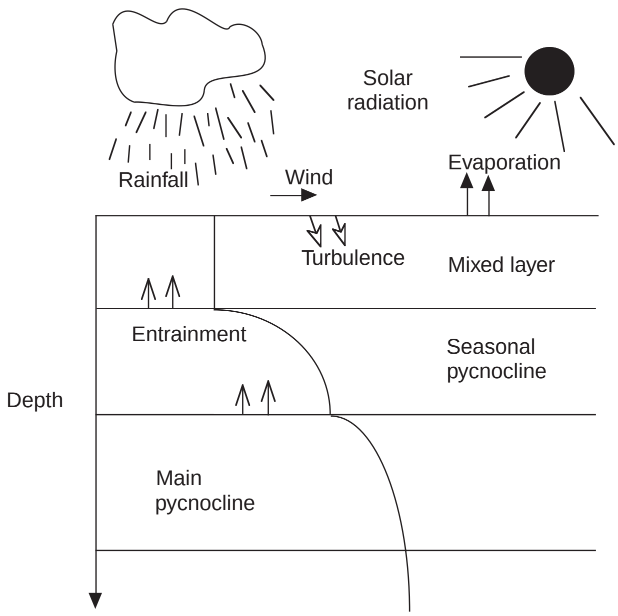

Figure 1 Conceptual diagram of the vertical structure in the surface layer, and the forcing and physics that govern its existence. The depth of the mixed layer, the seasonal pycnocline, and the main pycnocline are indicated.

overturning of water in the mixed layer by waves, and also by the entrainment of cooler water through the bottom of the mixed layer. Wind forcing also results in advection by upper ocean currents that can change the properties and thus the vertical structure of the mixed layer. Surface buoyancy forcing is due to heat and fresh water fluxed across the air–sea interface. Cooling and evaporation induce convective mixing and overturning, whereas heating and rainfall cause the mixed layer to restratify in depth and display alternate levels of greater and lesser vertical property gradients. Thus, if strong enough, the wind and buoyancy fluxes can generate sufficient turbulence so that the upper portion of the surface layer has a thick, homogeneous (low vertical gradient or stratification), well-mixed layer in temperature, salinity, and density.

Variations in the strength and relative contributions of the atmospheric forcing can cause substantial variability in the water properties and thickness of the upper mixed layer. Large temporal variation can occur on daily and seasonal timescales due to changes in the solar radiation. For example, during the daily cycle the sun heats the ocean causing the upper surface to become increasingly warm and weakly stratified. The 'classic' vertically uniform mixed layer may not be present in the upper ocean surface layer as depicted in **Figure 1**. As the sun sets, the surface waters are cooled and sink, generating turbulent convection that causes entrainment of water from below, and mixing that produces the vertically well-mixed layer. Temporal variability and the physics that govern this variability are covered elsewhere (*see* **Upper Ocean Time and Space Variability** and **Wind and Buoyancy-forced Upper Ocean**).

Furthermore, the mixed layer structure can exhibit large horizontal variations. For example, the large meridional differences in solar radiation result in mixed layers that generally increase in depth from the equator to the poles. Even in the east–west direction, boundary currents and differential surface forcing can result in mixed layers of different structures, although differences in the annual variations of temperature along a given latitude will generally be small. Spatial variability in the structure of the mixed layer is discussed elsewhere (*see* **Upper Ocean Mean Horizontal Structure**).

Separating the upper mixed layer from the deeper ocean is a region typically characterized by substantial vertical gradients in water properties. In temperature, this highly stratified vertical zone is referred to as the thermocline, and in salinity it is the halocline. Although the thermocline and the halocline may not always exactly coincide in their depth range, one or the other property will control the density structure to form the 'main pycnocline'. Because more often than not in the tropics and midlatitudes temperature is the controlling factor of density, the main pycnocline is also referred to as the 'permanent thermocline'. In midlatitudes during summer, surface heating from the sun can cause a shallow seasonal thermocline (pycnocline), that connects the upper mixed layer to the permanent thermocline (see **Figure 1**). Similarly, in the subpolar regions the seasonal summer input of fresh water at the surface through rainfall, rivers or ice melt, can result in a seasonal halocline (pycnocline) separating the fresh surface from the deeper saltier waters. Whereas the seasonal pycnocline disappears every winter, the permanent pycnocline is always present in these areas.

To maintain stability in the water column, lighter (less dense) water must lie above heavier (denser) water. It follows then, that the pycnocline is a region where density increases rapidly with depth. If the vertical density gradient is very strong, the turbulence within the upper mixed layer induced by the air–sea exchanges of wind and heat, cannot overcome the great stability of the pycnocline to penetrate into the deeper ocean. The base of the main pycnocline marks the depth limit of the upper ocean surface layer. Beneath this depth, the water has not seen the surface for a very long time, and above it, the stability of the main pycnocline acts as a barrier against turbulent mixing processes.

In some polar regions, particularly in the far North and South Atlantic, no permanent thermocline exists. The isothermal water column suggests that the cold, dense waters are virtually continuously sinking to great depths. No stable permanent pycnocline or thermocline exists as a barrier to the vertical passage of the surface water properties that extend to the bottom. In some cases, such as along the shelf in Antarctica's Weddell Sea in the South Atlantic, salinity can also play a role in dense water formation. When ice forms from the sea water in this region, it consists primarily of fresh water, and leaves behind a more saline and thus denser surface water that must also sink. The vertical flow of the dense waters in the polar regions is the source of the world's deep and bottom waters that then slowly mix and spread horizontally via the large-scale thermohaline ocean circulation to fill the deep ocean basins.

In fact, the thermohaline circulation also plays an important role in maintaining the permanent thermocline at a relatively constant depth in the low and middle latitudes. Despite the fact that the pycnocline is extremely stable, it might be assumed that on some timescale it could be eroded away through mixing of water above and below it. Humboldt recognized early in the nineteenth century that ocean circulation must help maintain the low temperatures of the deeper oceans; the equatorward movement of the cold deep and bottom water masses are continually renewed through sinking (or 'convection') in the polar region. However, it was not until the mid-twentieth century, that Stommel suggested that there was also a slow but continual upward movement of this cool water to balance the downward diffusion of heat from the surface. It is this balance, that occurs over very small space and timescales, that sustains the permanent thermocline observed at middle and low latitudes. Thus, the vertical structure of the upper ocean helps us to understand not only the wind- and thermohaline-forced ocean circulation, but also the response between the coupled air–sea system and the deeper ocean on a global scale.

Definitions

Surface Layer Depth

Conceptually the surface layer includes the mixed layer, where active air–sea exchanges are occurring, plus those waters below the mixed layer and above the base of the permanent thermocline that have been exposed to the atmosphere in their recent history. Note the important detail that the surface layer includes the mixed layer, a fact that has often been blurred in the criteria used to determine their respective depth levels. A satisfactory depth criterion for the surface layer should thus include all the major features of the upper ocean surface layer described above and illustrated in **Figure 1**. Further, the surface layer depth criterion should be applicable to all geographic regimes, and include those waters that have recently been in contact with the atmosphere, at least on timescales of up to a year. Finally, the definition should preferably be based on readily measurable properties such as temperature, salinity, or density.

Ideally then, we could specify the surface layer to be the depth where, for instance, the temperature is equal to the previous winter's minimum SST. However in practice, this surface layer definition would vary temporally, making it difficult to decipher year-to-year variability. Oceanographers therefore generally prefer a static criterion, and thus modify the definition to be the depth where the temperature is equal to the coldest SST ever observed using the historical data available at a particular geographic location. This definition is analogous to a local 'ventilation' depth: the deepest surface to which recent atmospheric influence has been felt at least over the timescale of the available historical data.

Note that the definition suggested for the surface layer is primarily one-dimensional involving only the temperature and salinity information from a given location. Lateral advective effects have not been included. The roles of velocity and shear, and three-dimensional processes in the surface layer structure may be important on occasions. However, their roles are harder to quantify and have not, as yet, been adequately incorporated into a working definition for the depth of the surface layer.

Mixed-layer Depth

The mixed layer is the upper portion of the surface layer where active air–sea exchanges generate surface turbulence which causes the water to mix and become vertically uniform in temperature and salinity, and thus density. Obviously direct measurement of the upper layer turbulence would provide the most accurate data for determining the mixed layer or 'mixing' depth, as dissipation levels typically decrease by an order of magnitude below the active convective depth. However, turbulence measurements can be problematic and are not widespread at present. For this reason, as in the surface-layer depth criterion, definitions of the mixed-layer depth are most commonly based on temperature, salinity, or density. The mixed-layer depth must define the depth of the transition from a homogeneous upper layer to the stratified layer of the pycnocline.

Two definitions of mixed-layer depth are widespread in the literature. The first determines the depth where a critical temperature or density gradient corresponding to the top of the maximum property gradient (i.e., the thermocline or pycnocline) is exceeded, and the second determines a net temperature or density change from the surface isotherm or isopycnal. The critical gradient criteria range between 0.02 and 0.05°C m^{-1} in temperature, and 0.005 and 0.015 σ_t units m^{-1} in density. This criterion may be sensitive to the vertical depth interval over which the gradient is calculated. In the second definition, common values used for the net change criterion are 0.5–1°C in temperature from the surface isotherm, or 0.125 σ_t units from the surface isopycnal. Because of the different dynamical processes associated with the molecular skin SST, oceanographers generally prefer the readily determined bulk SST estimate as the surface reference temperature. Ranges used in the temperature and density values used in both mixed-layer depth definitions will distinguish weakly stratified regions from unstratified. Another form of the net change criterion used to define the mixed-layer depth (mld), determines the depth at which,

$$\rho(z = \mathrm{mld}) = \rho(z = 0) - \delta T\, \mathrm{d}\rho/\mathrm{d}T \qquad [1]$$

where $\rho(z = 0)$ is the surface density, δT the net change in temperature from the surface (e.g. 0.5°–1°C), and $\mathrm{d}\rho/\mathrm{d}T$ is the thermal expansion coefficient calculated from the equation of state for sea water using surface temperature and salinity values. This criterion thus determines the depth at which density is greater than the surface density by an amount equivalent to the δT temperature change. Through the use of density and the thermal expansion coefficient, this definition has the advantage of revealing mixed layers where salinity stratification may be important. Criteria based on salinity changes, although inherent in the density criterion, are not evident in the literature as typically heat fluxes are large compared to freshwater fluxes, and the gravitational stability of the water column is typically controlled by the temperature stratification. In addition, subsurface salinity observations are not as regularly available as temperature.

To illustrate the differences between the mixed-layer depth criterion, **Figure 2(A)** shows the mixed-layer depth from an expendable conductivity–temperature–depth (XCTD: *see* **Expendable Sensors**) profile, using the net temperature (0.5°C) and density (0.125) σ_t change criteria, the gradient density criterion (0.01 m^{-1}), and eqn [1]. In this particular case, there is little difference between the mixed-layer depth determined from any method or property. However, **Figure 2(B)** shows an XCTD cast from the western Pacific Ocean, and the strong salinity halocline that defines the bottom of the upper mixed layer is only correctly identified using the density-defined criteria.

Finally, to illustrate the distinction between the surface layer and the upper mixed layer, **Figure 3(A)** shows a temperature section of the upper 300 m from Auckland to Seattle during April 1996. The corresponding temperature stratification (i.e., the vertical temperature gradient) is shown in **Figure 3(B)**. The surface layer, determined as the depth of the climatological minimum SST isotherm, and also the mixed-layer depth from a 1°C net temperature change from the surface (i.e., SST − 1°C) are indicated on both panels. This cross-equatorial north–south section also serves to illustrate the seasonal differences expected in the mixed layer. In the Southern Hemisphere in early fall the net temperature mixed-layer depth criterion picks out the top of the remaining seasonal thermocline, as depicted by the increase in temperature stratification in **Figure 3(B)**. The mixed-layer depth criterion therefore excludes information about the depth of the prior winter local wind stirring or heat exchange at the air–sea surface, that has been successfully captured in the surface layer using the historical minimum SST criterion. In the Northern Hemisphere tropical regions where there is little seasonal cycle, the surface layer and the mixed-layer criteria are nearly coincident. The depth of the mixed-layer and the surface layer extend down to the main thermocline. Finally in the early-spring northern latitudes, the mixed-layer criterion again mainly picks out the upper layer of increased stratification that was likely caused through early seasonal surface heating. The surface layer definition lies deeper in the water column within the main thermocline, and below a second layer of relatively low stratification. The deeper, weakly stratified region indicates the presence of fossil layers, which are defined in the next section.

Variability in Upper Ocean Vertical Structures

Fossil Layers

Fossil layers are nearly isothermal layers that occur below the upper well-mixed layer, and are separated by a well-stratified layer (see **Figure 3(B)**, 31°–37°N).

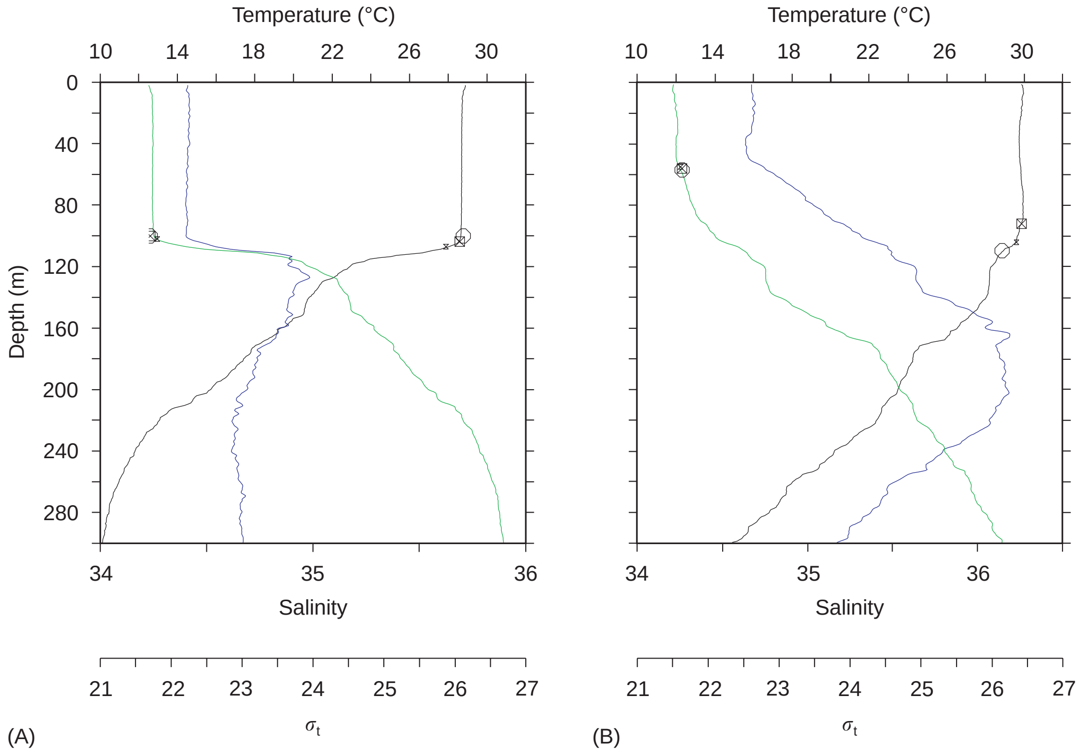

Figure 2 Temperature (black line), salinity (blue line) and density (green line) during March 1995 from expendable conductivity–temperature–depth profiles in the Pacific Ocean at (A) 6.9°N, 173.2°W and (B) 6°S, 166°W. In temperature, mixed-layer depth is calculated using criteria of a net temperature change of 0.5°C (crossed-box) and 1°C (circle) from the sea surface; and a temperature gradient criteria of 0.01°C m^{-1} (small cross). In density, mixed-layer depth is determined using criteria of a net density change of 0.125 σ_t units from the surface isopycnal (crossed box); a density gradient of 0.01 σ_t units m^{-1} (circle) and the thermal expansion method of eqn [1] (cross). Note the barrier layer defined as the difference between the deeper isothermal layer and the shallow density-defined mixed layer in (B).

The fact that these layers are warmer than the local minimum SST, defining the surface layer depth, indicates that they have at some time been subject to local surface forcing. The solar heating and reduced wind stirring of spring can cause the upper layer to become thermally restratified. The newly formed upper mixed layer of light, warm water is separated from the older, deeper winter mixed layer by a well-stratified thermocline. The fairly stable waters in this seasonal thermocline may isolate the lower isothermal layer and prevent further modification of its properties, so that this layer retains the water characteristics of its winter formation period and becomes 'fossilized'. Hence, fossil layers tend to form in regions with significant seasonal heating, a large annual range in wind stress, and deep winter mixed layers. These conditions can be found at the poleward edges of the subtropical gyres.

In the north-east Pacific Ocean off California and in the south-west Pacific Ocean near New Zealand, particularly deep and thick fossil layers have been associated with the formation of Subtropical Mode Waters. As with the fossil layers, the mode waters are characterized by low vertical gradients in temperature and density. The isothermal layer or thermostad of winter water trapped in the fossilized layers may be subducted into the permanent thermocline through the action of Ekman pumping, in response to a convergence in the wind field. The mode waters are then transported, retaining their characteristic thermostad, with flow in the subtropical gyre.

However, not all fossil layers are associated with the formation regions of mode waters. Shallow fossil layers have been observed where there are strong diurnal cycles, such as in the western equatorial Pacific Ocean. Here, the same alternating processes of heating/cooling and wind mixing as found in the mode water formation regions, cause the fossil layers to form. Fossil layers have also been observed around areas of abrupt topography, such as along island chains where strong currents are found. In this case, the fossil layers are probably formed by the advection of water with properties different from those found in the upper mixed layer.

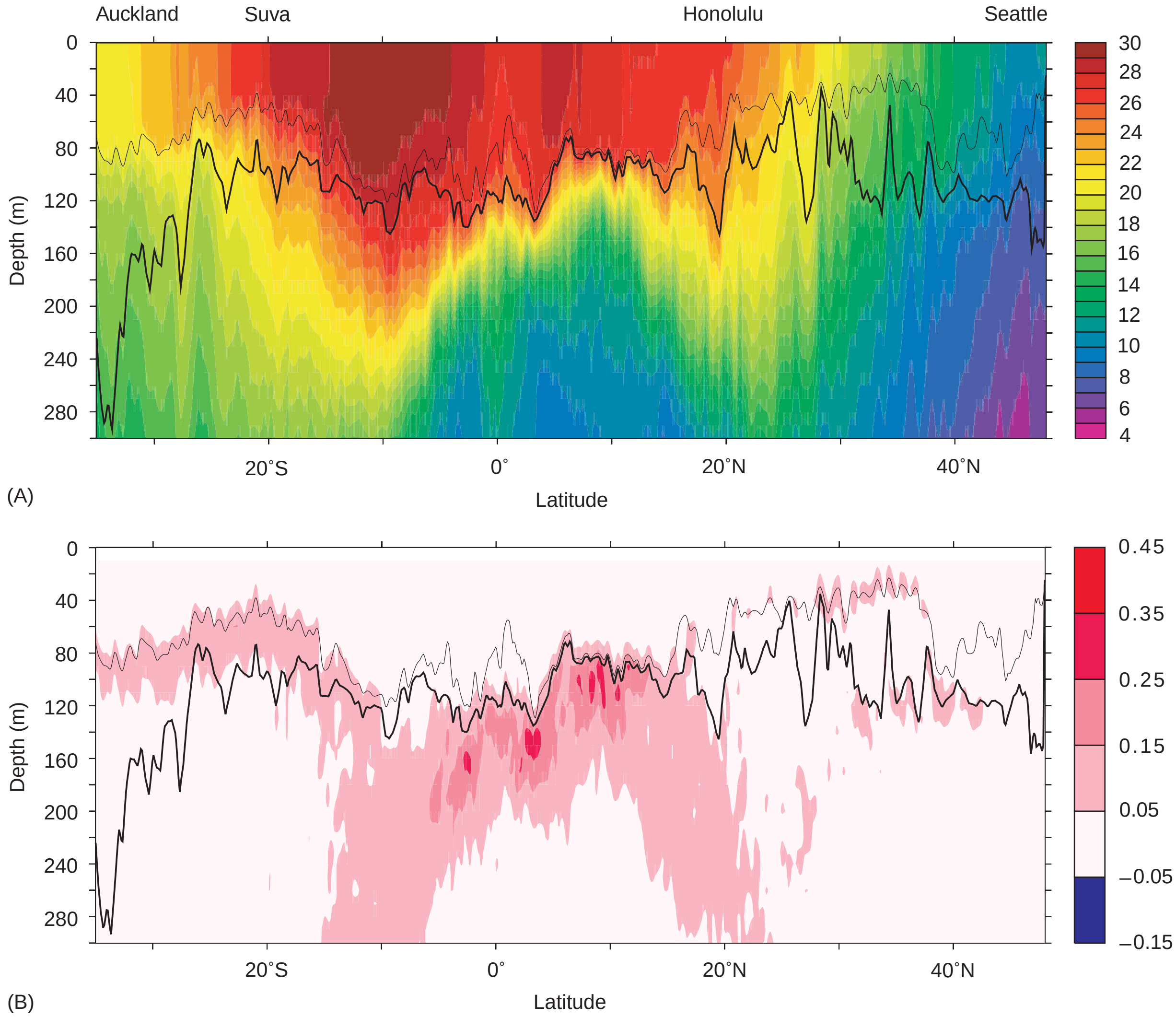

Figure 3 (A) The temperature section from expendable bathythermograph data collected along a transect from Auckland, New Zealand to Seattle in April 1996, and (B) the corresponding temperature gradient with depth. The heavy line indicates the depth of the surface layer, according to the depth of the coldest sea surface temperature measured at each location. The light line indicates the depth of the mixed layer according to the (SST – 1°C) criterion.

Barrier Layers

In some regions the freshwater flux can dominate the mixed-layer thermodynamics. This is evident in the tropics where heavy precipitation can cause a surface-trapped freshwater pool that forms a shallower mixed layer within a deeper nearly isothermal layer. The region between the shallower density defined well-mixed layer and the deeper isothermal layer (**Figure 2B**), is referred to as a salinity-stratified barrier layer.

The barrier layer may have important implications on the heat balance within the surface layer because, as the name suggests, it effectively limits communication between the ocean mixed layer and the deeper permanent thermocline. Even if under light wind conditions water is entrained from below into the mixed layer, it will have the same temperature as the water in this upper layer. Thus, there is no heat flux through the bottom of the mixed layer and other sinks must come into play to balance the solar warming that is confined to the surface, or more likely, the barrier layer is transient in nature.

Inversions

Occasionally temperature stratification within the surface layer can be inverted (i.e., cool water lies above warmer water). The temperature inversion can be maintained in a stable water column since it is density-compensated by a corresponding salinity increase with depth throughout the inversion layer. Inversions are a ubiquitous feature in the vertical structure of the surface layer from the equator to subpolar latitudes, although their shape and formation mechanisms may differ.

Inversions that form in response to a change in the seasonal heating at the surface are most commonly found in the subpolar regions. They can form when the relatively warmer surface water of summer is trapped by the cooler, fresher conditions that exist during winter. The vertical structure of the surface layer is described by a well-mixed upper layer in temperature, salinity, and density, lying above the inversion layer that contains the halocline and subsequent pycnocline (**Figure 4A**). Conversely, during summer the weak subpolar solar heating can trap the very cold surface waters of winter, sandwiching them between the warmer surface and deeper layers. In this case, the vertical structure of the surface layer consists of a temperature minimum layer below the warm stratified surface layer, and above the relatively warmer deeper layer (**Figure 4B**). The density-defined mixed layer occurs above the temperature minimum. With continual but slow summer heating, the cold water found in this inversion layer slowly mixes with the warmer water masses above and below, and erodes away.

Inversions can also form through horizontal advection of water with different properties or water-mass interleaving. For example, in the tropics where there may be velocity shear between opposing currents, inversions are typically characterized as small abrupt features, often only 10 m thick, found at the base of a well-mixed upper layer, and within the top of the halocline and pycnocline (**Figure 4C**). Just west of San Francisco (130°–140°W), the low temperature and salinity properties of the Subantarctic Water Mass found in the California Current transition toward the higher-salinity water masses formed in the evaporative-regime of the mid-subtropical gyre. The interleaving of the various water masses results in inversions that are quite different in structure from those observed in the tropical Pacific or the subpolar regions (**Figure 4D**). The surface layer vertical structure is further complicated by frequent energetic eddies and meanders that perturb the flow and have their own distinctive water properties. In the transition zone, the inversions can be thick, and occur well within the pycnocline and not at the base of the mixed layer (**Figure 4D**). Typically within the broad region of the inversion, there are sharp gradients in temperature and salinity, both horizontally and vertically, that are characteristic of water-mass interleaving from the advective penetrations of the currents and eddies.

Other Properties of the Upper Ocean Vertical Structure

Other water properties, such as dissolved oxygen and nutrients (e.g., phosphates, silica, and nitrates), can also vary in structure in the upper ocean surface layer. These properties are considered to be nonconservative, that is, their distribution in the water column may change as they are produced or consumed by marine organisms. Thus, although they are of great importance to the marine biology, their value in defining the physical structure of the upper ocean surface layer must be viewed with caution. In addition, until recently these properties were not routinely measured on hydrographic cruises. Nonetheless, the dissolved oxygen saturation of the upper ocean in particular, has been a useful property for determining the depth of penetration of air–sea exchanges, and also for tracing water masses. For example, in the far North Pacific Ocean, it has been suggested that the degree of saturation of the dissolved oxygen concentration may be a better indicator than temperature or density for determining the surface-layer depth of convective events. During summer, the upper layer may be restratified in temperature and salinity through local warming or freshening at the surface, or through the horizontal advection of less dense waters. However, these upper level processes may not erode the high-oxygen saturation signature of the deeper winter convection. Thus the deep high-oxygen saturation level provides a clear record of the depth of convective penetration from the air–sea exchange of the previous winter, and a unique signal for defining the true depth of the surface layer.

Conclusions

The vertical structure of the upper surface layer can be characterized at its simplest as having a near-surface mixed layer, below which there may exist a seasonal thermocline, where temperature changes relatively rapidly, connected to the permanent thermocline or main pycnocline. The vertical structure is primarily defined by stratification in the water properties of temperature, salinity, and density, although in some regions oxygen saturation and nutrient distribution can play an important biochemical role. The vertical structure of the surface layer can be complex and variable. There exist distinct variations

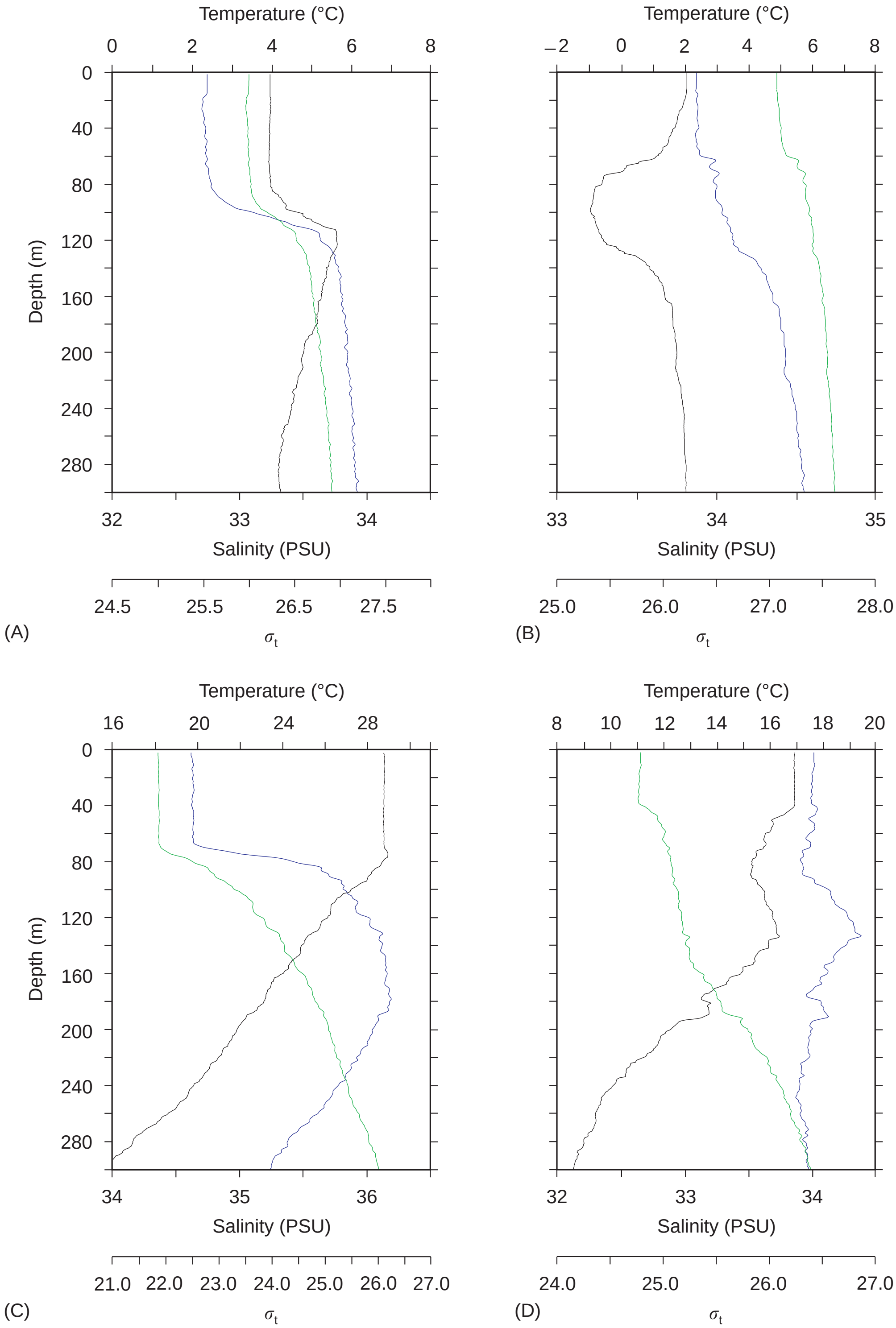

Figure 4 Temperature (black line), salinity (blue line) and density (σ_t, green line) from expendable conductivity–temperature–density casts at (A) 58.2°N, 147.3°W in March 1996, (B) 61°S, 63.9°W in January 1997, (C) 11.9°S, 176.1°W in August 1998, and (D) 33.5°N, 134.6°W in May 1995. Note the presence of temperature inversions at the base of the mixed layer in all casts.

in the forms and thickness of the upper-layer structure both in time and in space, through transient variations in the air–sea forcing from winds, heat, and fresh water that cause the turbulent mixing of the upper ocean. Understanding the variation in the upper ocean vertical structure is crucial for understanding the coupled air–sea climate system, and the storage of the heat and fresh water that is ultimately redistributed throughout the world oceans by the general circulation.

See also

Air–Sea Gas Exchange. Bottom Water Formation. Deep Convection. Ekman Transport and Pumping. Expendable Sensors. Heat and Momentum Fluxes at the Sea Surface. Ocean Subduction. Open Ocean Convection. Penetrating Shortwave Radiation. Thermohaline Circulation. Upper Ocean Heat and Freshwater Budgets. Upper Ocean Mean Horizontal Structure. Upper Ocean Mixing Processes. Upper Ocean Time and Space Variability. Water Types and Water Masses. Wind Driven Circulation. Wind and Buoyancy-forced Upper Ocean.

Further reading

Kraus EB and Businger JA (1994) *Atmosphere–Ocean Interaction*, Oxford Monographs on Geology and Geophysics, 2nd edn. New York: Oxford University Press.

Philips OM (1977) *The Dynamics of the Upper Ocean*, 2nd edn. London and New York: Cambridge University press.

Reid JL (1982) On the use of dissolved oxygen concentration as an indicator of winter convection. *Naval Research Reviews* 3: 28–39.

Warren B and Wunsch W (1981) *Evolution of Physical Oceanography: Scientific Surveys in Honor of Henry Stommel*. Cambridge, MA: MIT Press.

Tomczak M and Godfrey JS (1994) *Regional Oceanography: An Introduction*. Oxford: Pergamon Press.

UPWELLING ECOSYSTEMS

R. T. Barber, Duke University Marine Laboratory, Beaufort, NC, USA

doi:10.1006/rwos.2001.0295

Introduction

An ecosystem is a natural unit in which physical and biological processes interact to organize the flow of energy, mass, and information. The result of this self-organizing activity is that each kind of ecosystem has a characteristic trophic structure and material cycle, some degree of internal homogeneity, objectively definable boundaries, and predictable patterns of seasonality. Oceanic ecosystems are those ecosystems that exist in the open ocean independently of solid substrates; for example, oceanic ecosystems are fundamentally distinct from coral or intertidal ecosystems.

Upwelling ecosystems are those that occupy regions of the ocean where there is a persistent upward motion of sea water that transports subsurface water with increased inorganic plant nutrients into the sunlit surface layer. The upwelling water is not only rich in nutrients, but also frequently cooler than the surface water it replaces; this results in a variety of atmospheric changes, such as coastal deserts or arid zones. The increased nutrient supply and favorable light regime of upwelling ecosystems, however, distinguish them from other oceanic ecosystems and generate characteristic food webs that are both quantitatively and qualitatively different from those of other oceanic ecosystems.

For persistent upwelling to take place it is necessary for the surface layer to be displaced laterally in a process physical oceanographers call divergence and then for subsurface water to flow upward to replace the displaced water. The physical concept of upwelling is simple in principle but, as with many ocean processes, it becomes surprisingly complex when real examples are studied. To begin with, there are two fundamental kinds of upwelling ecosystems: coastal and oceanic. They differ in the nature of their divergence. In coastal upwelling, the surface layer diverges from the coastline and flows offshore in a shallow layer; subsurface water flows inshore toward the coast, up to the surface layer, then offshore in the surface divergence. In contrast, oceanic upwelling, which occurs in many regions of the ocean, depends on the divergence of one surface layer of water from another. One such oceanic divergence is created when an increasing gradient in wind strength forces one surface layer to move faster, thereby leaving behind, or diverging from, another surface layer. Major regions of this kind of oceanic upwelling are found in high latitudes in the Subpolar gyres of the Northern Hemisphere and the Antarctic divergence in the Southern Ocean. The food webs of polar upwelling ecosystems are described elsewhere in the Encyclopedia and this article will focus on coastal and equatorial upwelling ecosystems that occur in low and mid-latitude regions of the world's oceans.

The physical boundary organizing oceanic divergence in equatorial upwelling is the Coriolis force, which changes sign at the equator, causing the easterly Trade Winds to force a northward divergence