Ocean Mixed Layer

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Introduction

The ocean mixed layer (OML), the ocean region adjacent to the air–sea interface, is typically tens of meters deep, and due to the fact that it is well mixed, the temperature and salinity (and therefore the density) are fairly uniform. The rapidly changing regions below these uniform regions of temperature, salinity, and density are called the thermocline, halocline, and pycnocline, respectively. The mixing is primarily shear-driven, since the wind stress at the surface is the primary mixing agent, although at night significant convective mixing driven by the heat loss to the atmosphere takes place. The OML is heated near the surface by both short-wave (SW) and long-wave (LW) radiative fluxes, and deeper in the water column from solar radiation in the visible part of the spectrum penetrating into the OML. This solar heating produces a diurnal cycle that varies in importance and magnitude at different latitudes. The cooling, however, is driven from heat and evaporative losses at the surface. Seasonal variation of the OML due to radiative heating is also important, although the importance depends on the latitude.

The OML mediates the exchange of mass, momentum, energy, and heat between the atmosphere and the ocean and hence plays a central role in long-term climate and weather. Because of the high heat capacity of water (2.5 m of the upper ocean has the same heat capacity as the entire troposphere), and because the oceans compose over two-thirds of the surface of the globe, most of the solar heating on Earth passes through the OML. Oceans are heat reservoirs, gaining heat during spring and summer and losing it slowly during fall and winter, and therefore act like a flywheel in matters related to weather on time scales of weeks and longer.

The OML also plays an important role in the oceanic food chain. Primary production by phytoplankton is the first link in this chain. The need for an energy source in producing biomass restricts primary production to the upper few tens of meters (the euphotic or photic zone), in which the solar insolation is strong enough to assist carbon fixation. The mixing at the base of the OML is also crucial to biological productivity. The OML is normally nutrient-poor, and it is the injection of nutrients from the nutrient-rich waters below the seasonal thermocline that permits higher levels of primary productivity. In fact, it is the upwelling regions (which compose just a few percent of the world's oceans), where nutrient-rich waters are forced into the OML and brought into the photic zone, that provide most of the fish catch around the world.

Biological productivity is important from a climatic point of view over time scales of decades or more. Carbon fixing constitutes a biological pathway for removing some of the anthropogenic CO_2 introduced into the atmosphere. There also exists an inorganic pathway, since there is a significant uptake of CO_2 in the cold subpolar oceans, some of which are also regions of deep and intermediate water formation. It is likely that the ocean acts as an important CO_2 sink on the globe and accounts for a significant fraction of the 'missing' anthropogenic CO_2 input to the atmosphere. However, quantification of the magnitude of this sink requires accurate OML models coupled to accurate ecosystem and air–sea transfer models.

Finally, the OML constitutes the first link in the chain of oceanic pollution. Most of the pollution in the global oceans takes place in the coastal oceans through the OML, and therefore the fate of any pollutants accidentally or intentionally deposited in the OML depends on the mixing and dispersion in the OML.

Characteristics

An OML can be divided into four parts: the very thin but important molecular sublayer, a few millimeters thick; the wave sublayer, normally 2–6 m thick; the main bulk of the OML, 10–40 m thick; and the entrainment sublayer of about 5–10 m thickness. In deep convective OMLs, where the mixed layer depth is a few hundred meters or more, the fractions of the wave and entrainment sublayers are small. In a shallow diurnal OML, a few meters thick, the wave sublayer can be a large fraction. An active gravity wave field can damp out the diurnal modulation of sea surface temperature (SST) by wave-driven mixing through Langmuir cells or wave breaking processes.

The active turbulent mixed layer in the upper ocean is usually bounded below by a strong buoyancy interface, in the form of a layer with either a sharp decrease in temperature (seasonal thermocline) or a sharp increase in salinity (halocline) (or both). In either case, this layer (called a pycnocline) is stably stratified, and here turbulence is damped by buoyancy forces.

The transition region from active turbulent mixing to mostly quiescent layers below can be called a turbucline, in analogy with the thermocline. Normally, the turbucline coincides with the seasonal thermocline or halocline, but not necessarily both. During high-precipitation events, a shallow brackish layer can form and the halocline and turbucline are at similar depth but the thermocline is much deeper. In the tropical Western Pacific, a similar situation can exist, leading to the so-called barrier layer that plays an important role in the transfer of heat from the ocean to the atmosphere in the tropics, by acting as a barrier to mixed layer deepening and entrainment of waters below the halocline.

An OML is mixed from both the top and the bottom. At the top, it is the winds, waves, and buoyancy fluxes that stir the fluid. At the bottom, it is the entrainment driven by large turbulent eddies in the OML that mixes the denser fluid from below into the OML. Wind-driven current in the OML also causes strong shear at the base of the mixed layer; shear instability ensues, inducing Kelvin–Helmholtz billows, which thicken the buoyancy interface and hence decrease its resistance to erosion by turbulent eddies. In deep OMLs, it is these mechanisms at the bottom that are responsible for a majority of the deepening of the OML. In shallow OMLs, the surface-stirring processes due to gravity wave breaking and cellular motions are also important. Note that turbulent erosion tends to sharpen the pycnocline, while K–H billows tend to make it more diffuse.

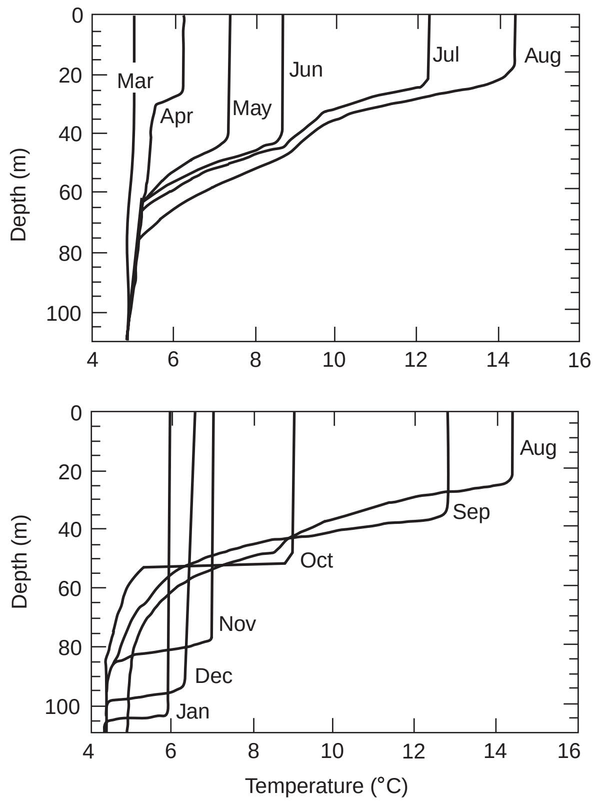

Figure 1 Seasonal evolution of temperature in the midlatitude upper ocean. Shallow, warm mixed layers during spring–summer alternate with cold, deep ones during fall–winter. (Kantha and Clayson, 2000.)

Perhaps the most salient aspect of the OML in midlatitudes is its diurnal and seasonal variability. **Figure 1** shows the typical seasonal cycle in midlatitudes. This seasonal variability in OML depth and temperature, and hence the heat content of the OML, is a prime factor in the air–sea exchange at these latitudes. The onset of spring warming restratifies the water column, and its depth stays roughly constant once the shallow spring–summertime thermocline forms. However, the formation period is heavily influenced by wind events at the time. Similarly, wind forcing controls the deepening of the OML at the onset of autumn cooling. During this time, the OML deepens episodically during intense storms that pass through the region, with significant assistance from cooling at the surface. Both sensible and latent heat fluxes are important in cooling the ocean at midlatitudes, whereas it is principally the evaporative losses that dominate the air–sea exchange at warmer low latitudes and sensible heat loss at colder high latitudes. Precipitation events also play a role in mixing. Wintertime cold air outbreaks along the east side of continents lead to rapid OML deepening, a large heat loss from the ocean, and cyclogenesis in the atmosphere. At higher latitudes, the OML deepens more because of penetrative convection, and in subpolar regions deep convection occurs. The OML structure in midlatitudes is affected by both salinity and temperature, whereas in subpolar and polar oceans salinity plays an overwhelmingly important role in mixing, and in the tropics the thermal structure in general predominates.

Diurnal variability affects the heat exchange on shorter time scales and may play a role over long time scales as well. The intensity of diurnal modulation of the OML depth and temperature depends on the season. Generally, the modulation is stronger if the solar insolation is strong and winds weak. This situation is typical of summer. Insolations as high as $1000\,\mathrm{W\,m^{-2}}$ are possible during clear summer days. This combined with very low winds gives rise to a diurnal modulation of as much as 2–4°C. Part of the heat built up in the OML during the day is lost by nocturnal cooling, which drives a vigorous convection and mixing in the water column that normally mixes some of the heat gained into the seasonal mixed layer.

A major factor in OML dynamics in the equatorial regions is the presence of strong background currents in the vicinity of the OML. The Equatorial Undercurrent in the Equatorial Pacific is a typical example. It exists at depths ranging from 50 to 200 m and is an eastward-flowing current that produces a strong vertical shear, which has a major influence on mixing in the upper water column. In contrast, in midlatitude oceans, the principal balance is between the Coriolis terms and the stress divergence, and the currents are not continuously accelerated by a steady wind; instead a steady state is reached and an Ekman-like spiral is produced.

In ice-covered oceans, the ice mediates the exchange of momentum between the atmosphere and the OML. The principal balance in ice is between the Coriolis force, the wind stress at the top, internal stresses in ice, and the shear stress on the ocean at the bottom. This force balance determines the stress available for mixing under ice. In addition, ice growth and melting cause buoyancy fluxes that affect the OML below the ice. Stirring by deep ice keels is an important factor. The ice cover tends to insulate high-latitude oceans from the cold atmosphere. However, it is not continuous, and even in the middle of winter there exist narrow openings in ice, called leads, through which a substantial fraction of wintertime heat loss to the atmosphere at high latitudes takes place. **Figure 2** shows the mixing processes prevalent under leads.

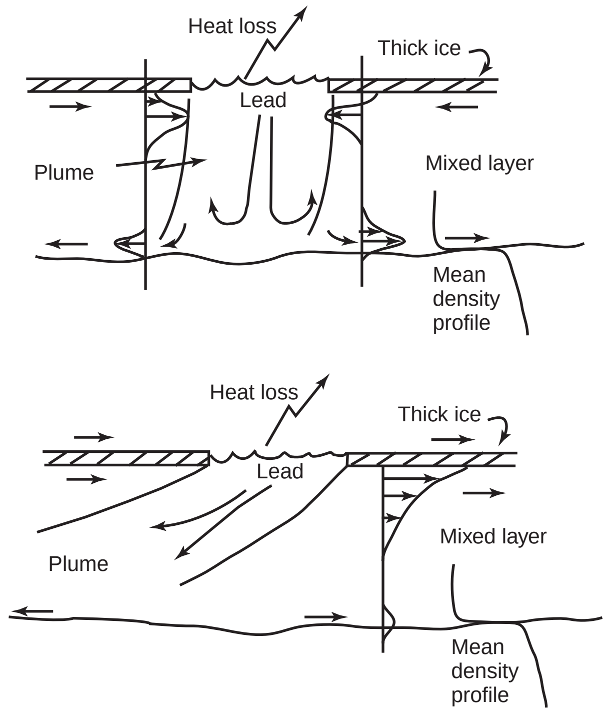

Figure 2 OML driven by freezing and salt extrusion underneath a refreezing lead. The top panel corresponds to stationary ice cover, when free convection dominates, and the bottom panel to the case when the ice is in motion, when forced convection dominates the mixing process. (Kantha and Clayson, 2000.)

There are striking similarities between the atmospheric boundary layer (ABL) over land and the OML. Under convective conditions, similar scaling laws hold in both turbulent layers. However, the most important difference between the OML and the ABL over land is the presence of surface waves at the air–sea interface that play an active role in its dynamics. The dynamical influence of the ground surface on the ABL is determined by its roughness and topography, which are invariant, whereas it is the effective roughness of the mobile sea surface that is constantly changing with the winds that is important in the OML.

Under neutral stratification, it is possible to find a region where the universal law of the wall scaling would apply: $q \approx u_*$, $l \approx z$ and $\varepsilon \approx z^{-1}$, $\nu_t \approx z$, where u_* is the frictional velocity (the square root of the ratio of the wind stress to the water density), q is the turbulence velocity scale, l its length scale, ε the dissipation rate of turbulence kinetic energy (TKE), ν_t the eddy viscosity, and z the depth. Here, the mean shear is proportional to u_* but inversely proportional to z, and therefore the mean velocity is proportional to the logarithm of the distance from the free surface (see the references listed under Further Reading). Indeed, this scaling can be found in the upper part of the OML, except close to the surface. Close to the surface, under strong wind conditions, modern measurements have found that the dissipation rate is one to two orders of magnitude larger than that given by the law of the wall (**Figure 3**). This near-surface elevated dissipation rate is due to the influence of surface waves and wave breaking. Wave breaking generates intermittent, shear-free turbulence somewhat akin to the turbulence generated by a stirring grid in a fluid. The turbulence intensity drops off sharply away from the source. Therefore, while the turbulence intensities are elevated above the usual levels during extensive wind-wave breaking, this turbulence is important only to a depth on the order of the amplitude of the breaking waves. Below these depths, the law of the wall can often be found once again. Wave breaking and associated turbulence are likely to be important for the dynamics of OMLs, especially shallow ones; because of the elevated near-surface dissipation rates, they may bring about a higher exchange of gas and heat across the air–sea interface. Were it not for the surface waves, the turbulence near the surface of an OML would behave roughly similar to that adjacent to a solid boundary, such as the ABL over land.

Solar Heating

The solar radiation incident on the ocean surface can be divided into three components: short wavelengths

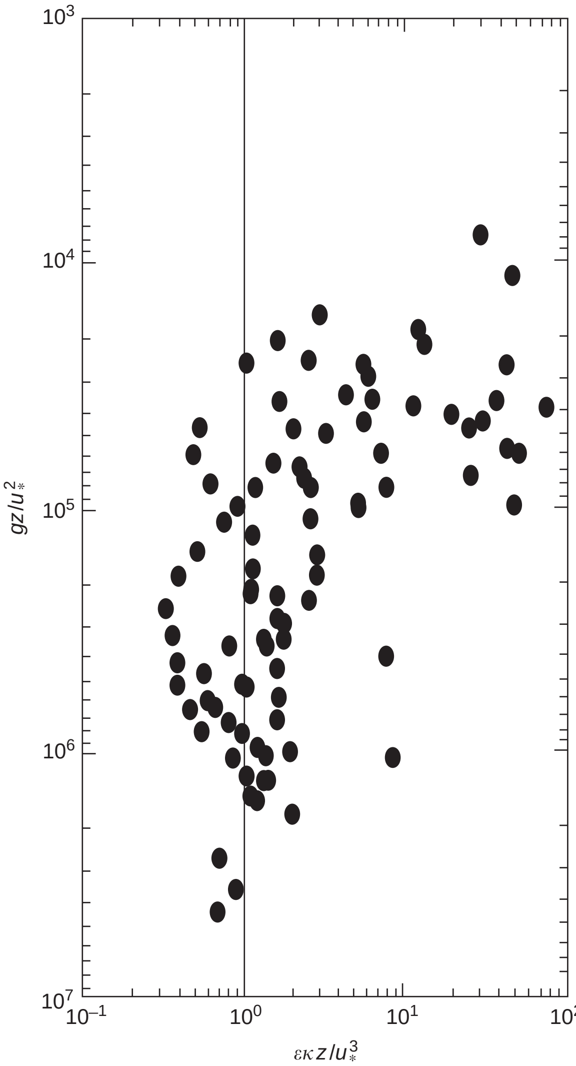

Figure 3 Measured dissipation rate of TKE in the near-surface layers of the OML. The dissipation rate is more than an order of magnitude higher than that expected from the classical law of the wall scaling. *g* is the gravitational acceleration and κ is the von Karman constant. (Kantha and Clayson, 2000.)

in the ultraviolet part of the spectrum (< 350 nm), the wavelengths available for photosynthesis (photosynthetically available radiation (PAR), 350–700 nm), and the infrared and near-infrared wavelengths (> 700 nm). The UV portion is roughly 2%, PAR 53%, and IR 45% of the total solar insolation. PAR coincides roughly with the visible portion of the spectrum and is the most important of the three portions for biological aspects of the upper ocean. Primary productivity and fixing of carbon by phytoplankton take place only in the euphotic zone, defined as the depth at which respiration and primary production balance, which is where PAR decreases to 1% of its surface value. The ultraviolet part is

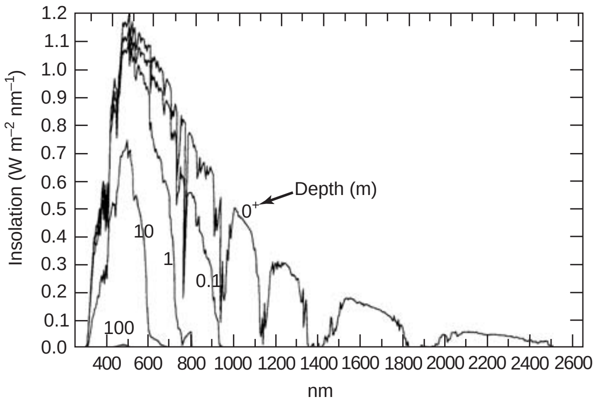

Figure 4 Solar insolation at the ocean surface and below. Note the rapid attenuation of nonvisible components with depth. Only the visible part remains below about 10 m. (Kantha and Clayson, 2000.)

important to the production of certain photochemicals such as carbonyl sulfide. **Figure 4** shows the spectrum of solar insolation at the top of the ocean and at various depths. Only the visible part remains below about 10 m.

Air–sea Fluxes

For air–sea exchange purposes, the momentum flux from the atmosphere to the oceans is the most important parameter, and therefore the transfer of momentum from winds to surface waves is usually relevant only in so far as it affects the net transfer to the ocean currents. Similarly, the water vapor and gas fluxes from the oceans to the atmosphere, which are usually net losses to the ocean, are also important. All these fluxes of heat, mass, momentum, and gases are determined by turbulent processes in the surface layers of the atmosphere and the ocean adjacent to the air–sea interface, with surface waves playing an important role by virtue of their ability to act as sinks of momentum, to determine the 'roughness' of the sea surface, and to disrupt the aqueous molecular sublayer responsible for transfer of scalar properties across the interface. At sufficiently high wind speeds, spray and droplets ejected into the atmosphere and air bubbles entrained into the ocean during wave breaking directly affect the water vapor and gas exchange between the two media.

The net fluxes of all scalars, including heat, must vanish at the interface (**Figure 5**).

$$SW_{\downarrow} + LW_{\downarrow} + H_{\mathrm{pr}} - SW_{\uparrow} - LW_{\uparrow} - H_{\mathrm{s}} - H_{\mathrm{L}} - SW_{\downarrow}^{\mathrm{P}} = 0 \qquad [1]$$

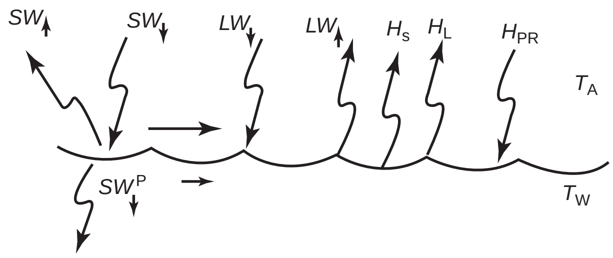

Figure 5 A sketch of the heat fluxes at the air–sea interface. (Kantha and Clayson, 2000.)

where $SW_{\uparrow} = \alpha SW_{\downarrow}$; α is the albedo of the ocean surface. There is thus a balance between the downwelling short-wave ($SW_{\downarrow}$) and long-wave ($LW_{\downarrow}$) radiative fluxes, and the upwelling $SW_{\uparrow}$ and $LW_{\uparrow}$ fluxes, the sensible (H_s) and latent (H_L) heat fluxes, the heat flux due to any precipitation (H_{pr}), and the solar radiative flux penetrating into the ocean ($SW_{\downarrow}^{P}$). The net salinity flux to the ocean is

$$F_{Spr} = (\dot{P}_{r,sn} - \dot{E})(-S_s) \qquad \dot{E} = H_L/L_E \qquad [2]$$

where $\dot{P}_{r,sn}$ is the precipitation (rain or snow) rate (m s^{-1}), $\dot{E}$ the evaporation rate, S_s the surface salinity, and L_E the latent heat of evaporation. Precipitation tends to suppress mixing because of the stabilizing effect of fresh water precipitated onto the salty water of the ocean. The net buoyancy flux due to precipitation is normally stabilizing, since the salinity effect overwhelms any thermal effect.

Momentum is a conserved quantity. Therefore, the momentum flux must also be in balance at the air–sea interface. This just means continuity of stresses at the interface, in particular, tangential stresses,

$$\tau_a + \tau_{pr} = \tau_w + \tau_{wv} \qquad [3]$$

where τ_a is the air-side stress (the shear stress applied by the atmosphere to the ocean), τ_w the water-side stress (negative of the drag exerted by the ocean on the atmosphere), τ_{wv} the momentum flux radiated out by propagating surface waves generated by the wind, and τ_{pr} the momentum flux due to precipitation. Enormous effort has gone into parameterizing τ_a in terms of the atmospheric variables and τ_{wv}, since they determine the value of τ_w. The momentum flux to the surface waves is a drag exerted on the atmosphere and is therefore important. It is especially important in the initial stages of development of the wave field (meaning short fetches or immediately following a change in the wind), since a considerable fraction of the momentum flux from the atmosphere goes then into generating the waves, with the remainder going directly into ocean currents. For a mature wave field, however, near-equilibrium conditions prevail and most of the momentum flux put into waves is immediately 'lost' and transferred to the currents, and τ_{wv} can be safely neglected. It is difficult in practice to compute τ_{wv} without a wave generation model, and it is a normal practice to ignore τ_{wv} and put $\tau_w = \tau_a$ in the absence of any precipitation.

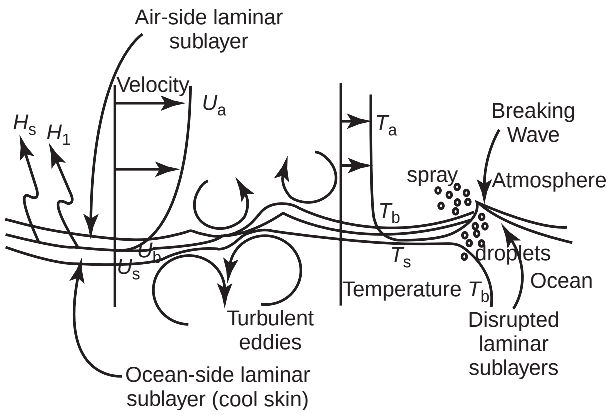

Figure 6 A sketch of the air–sea interface showing the molecular sublayers on the air and the water sides. The heat flux through the sublayer on the water side is responsible for the large temperature gradient there and hence the cool skin of the ocean. (Kantha and Clayson, 2000.)

Heating and cooling of the ocean surface occur across the skin of the ocean. Within this skin layer, which is on the order of a millimeter in thickness, there is usually a sharp drop in temperature of a few tenths of a degree Celsius (**Figure 6**). Exchanges of heat, momentum, and mass through this region are by molecular processes. This cool skin plays an important role in air–sea transfer processes, because of its influence on air–sea temperature and humidity differences. Transport of dissolved gases also occurs across a molecular sublayer of similar thickness to the thermal sublayer. Below this thin layer, turbulent processes dominate, driven by momentum and energy exchanges from the atmosphere to the ocean. The presence of the wave sublayer and a mobile interface, whose roughness as felt by both the ocean and the atmosphere is dynamically determined, provides the most important distinction between an atmospheric boundary layer (ABL) over land and that over the ocean. Consequently, surface layer similarity laws derived from the ABL over land are imprecise analogies when applied to the ABL over water and the OML. This is more so for the OML where the fraction of the OML affected directly by wave orbital velocities and hence wave dynamics can be large.

Langmuir Cells

Langmuir cells are organized counterrotating cells in the surface layer (**Figure 7**), with axes roughly aligned

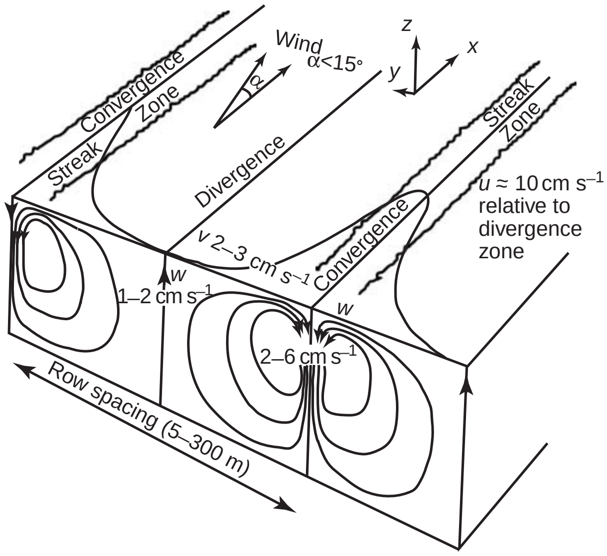

Figure 7 Langmuir cells and the associated velocities. Note the "windrows" and the strong near-surface vertical velocities associated with the convergence zones of a pair of counter rotating cells. Note the zones indicated as streak zones are the 'windrows'. α is the angle between the wind and the windrows, w is the vertical velocity, u and v are velocities in the x and y directions. (Kantha and Clayson, 2000.)

with the wind. Their presence is indicated by surface convergence at the boundary of counterrotating cells, where seaweed and flotsam accumulate. The convergence region is also made visible by whitecapping and bubble entrainment due to breaking of small-scale waves, resulting in parallel white lines aligned roughly with the wind and roughly uniformly spaced. Bubble clouds in water are also concentrated at surface convergence zones and are visible in side-scan sonar observations.

Organized motions in the OML such as those due to large eddies and Langmuir circulations are important to upper-ocean mixing and transport. Langmuir cells can be quite vigorous with downward vertical velocity immediately below the convergence zone as high as a few tens of cm centimeters per second. These motions are capable not only of injecting additional energy into the OML for mixing, but of also transporting floating particles such as phytoplankton deep into the OML. They are transient processes, however, and not easily quantified. Their existence depends on the presence of a surface wave field with the associated Stokes drift, a small steady Lagrangian residual current in the direction of surface wave propagation which decays exponentially with depth. An instability brought on by the vortex force term that appears in the momentum equations due to the interaction of the Stokes drift with the mean shear in the upper layers leads to the formation of Langmuir cells. Thus they are unique to the oceans since they result from a subtle interaction of the wind-driven turbulence and the Stokes current drift produced by surface gravity waves. Observational programs and advanced computer models such as large eddy simulations (LES) are helping us understand such large-scale features of the OML. Langmuir cells can have a dramatic effect on shallow diurnal mixed layers and can wipe out the strong diurnal peaks in the sea surface temperature that would otherwise be manifest when solar insolation is strong and winds are weak. Their effect on mixing in deep mixed layers can also be significant, even though the Stokes drift decays rapidly with depth.

The characteristic velocity scale for Langmuir circulation is

$$V_{\mathrm{L}} = [u_*^2 (ka)^2 \, C \cos\theta]^{1/3} \qquad [4]$$

where u_* is the friction velocity, k the wavenumber, C the phase velocity, a the amplitude of the surface waves, and θ the angle between wind stress and the direction of wave propagation. Clearly, the strength of the Langmuir cells depends on both the Stokes drift and wind stress, so that strong winds and small waves can have an influence similar to that of weak winds and large waves.

Deep Convection

There exist a few regions in the high-latitude oceans where convection is both deep and long-lasting. Under the present climatic conditions, in the Greenland Sea, the Labrador Sea, and the western Mediterranean Sea in the Northern Hemisphere, and in the Weddell Sea in the Southern Hemisphere, strong, prolonged wintertime cooling occurs at the surface, leading to deep convective layers that extend over most of the water column. Deep convection in the open ocean is the means by which the deep ocean is ventilated and its thermal structure maintained. The resulting meridional thermohaline circulation, and the poleward oceanic heat transport from the low latitudes to the mid–high latitudes associated with it, have a major influence on the climate at these latitudes. In these few deep convection regions, the stable stratification in the water column that normally isolates the abyss from the atmosphere is broken down violently by strong convective cooling at the surface. The associated timescales are much larger than the inertial period ($2\pi/f$, where $f = 2\Omega \sin\theta$ is the Coriolis parameter, Ω being the angular velocity of Earth and θ being the latitude) and hence deep convection occurs under the influence of Earth's rotation.

There are three phases of deep convection as follows:

(1) The preconditioning phase, where the prevailing large-scale circulation brings the weakly stratified deep water masses closer to the surface for the stratification to be gradually eroded by strong, sustained surface cooling during early winter. This phase is crucial to the whole process. Deep convection in the open ocean is found only in regions with cyclonic circulation that causes an upward doming of the isotherms. In the Labrador Sea, the cyclonic circulation is due to the West Greenland and Labrador currents hugging the continental slope. In the Mediterranean, the cyclonic Lions Gyre provides the preconditioning. Strong, sustained cooling is also essential to break down the stratification built up in the upper layers during the previous spring and summer. It is interesting that even stronger heat losses ($\sim 1000\,\mathrm{W\,m^{-2}}$) occur in the oceans during wintertime cold-air outbreaks off the east coasts of continents leading to strong cyclogenesis in the atmosphere, but not deep-water formation because of the brevity of the event. In the polar oceans, the air–sea temperature differences during off-ice wind conditions can reach 30–40°C, and if sustained long enough can lead to intermediate and deep water formation, examples being the Weddell Sea in the Antarctic, the Sea of Okhotsk, and the Arctic shelves.

(2) Eventually the stratification breaks down. Strong cooling events lasting several days with heat losses of $500\text{–}1000\,\mathrm{W\,m^{-2}}$, brought on by air–sea temperature differences of 8–12°C, and strong wind bursts are common. Deep convection ensues, with intense plumes a few kilometers in size reaching down to as deep as 2–3 km.

(3) The cooling weakens, and the well-mixed water mass in the convective 'chimney' spreads laterally, undergoing baroclinic instabilities in the process, and mixes with the ambient waters. Stratification is restored and the stage is set for the next cycle.

The most relevant parameters in deep convection are the buoyancy flux (due to both sensible and evaporative heat losses), B_0, that can reach values of $1\text{–}3 \times 10^{-7}\,\mathrm{m^2\,s^{-3}}$, the inertial frequency, f, the depth of the mixed layer, D, 1000–3000 m, and the Rossby radius of deformation, $a = c/f$ (where c is the internal gravity wave speed), typically a few kilometers. The Rossby radius is indicative of the horizontal scales of motion under the influence of ambient rotation. Under conditions where the rotational effects dominate, the relevant length and velocity scales are $l_{\mathrm{dc}} = (B_0/f^3)^{1/2}$ and $u_{\mathrm{dc}} = (B_0/f)^{1/2}$. The associated Rossby number, $Ro = u_{\mathrm{dc}}/(fl_{\mathrm{dc}})$, characterizing the relative importance of rotation is unity. The relevant Rayleigh number, a parameter of importance in thermal instability and free convection, is $Ra = B_0D^4/(\nu k^2)$, on the order of 10^{26}. Note that in the atmosphere, where rotational effects are not important, the relevant length scale is D, and the relevant velocity scale is the Deardorff scale, $w_* = (B_0D)^{1/3}$, indicative of the typical velocities in a convective ABL.

Numerical Models

Oceanic mixed layer models can be grouped into roughly two categories: bulk (or slab) models and diffusion models. Bulk models attempt to model the OML in an integral sense. The governing equations are integrated over the mixed layer so that the momentum and heat balance of the entire mixed layer, under the action of momentum and buoyancy fluxes at the ocean surface, can be considered. The major problem in bulk mixed layer modeling arises from the necessity to parameterize the advance and retreat of the OML under the action of surface momentum and buoyancy fluxes. The entrainment rate at the base of the OML, determined by turbulence processes, governs the deepening of the OML. This has been a subject of both laboratory and field experiments. It is also necessary to know the depth to which turbulence generated at the surface can penetrate under the action of a stabilizing buoyancy flux at the surface, as, for example, during a rainstorm or strong solar heating. Bulk models parameterize the entrainment (OML deepening) and 'detrainment' (OML retreat) in terms of surface fluxes of momentum and buoyancy, using well-known properties of turbulence in geophysical mixed layers and/or observational evidence.

Diffusion models parameterize turbulent mixing in the OML. Those based on higher moments of governing equations close the governing equations for turbulence quantities at some level by judicious modeling of the unknown higher moments and other terms. Once turbulence is thus quantified, it is a straightforward matter to deduce the mixing intensity. In second-moment closure models, the turbulence equations are closed at the second-moment level; conservation equations for turbulence Reynolds stresses, heat fluxes, and scalar variances are solved by modeling the unknown third-moment turbulence quantities and pressure–strain rate and pressure–scalar gradient covariance terms by appealing to physical intuition and/or observational evidence (and lately to large eddy simulations of turbulence). However, the complexity of this approach is at least an order of magnitude more than that of the simpler models cited above, since there is now a need to solve

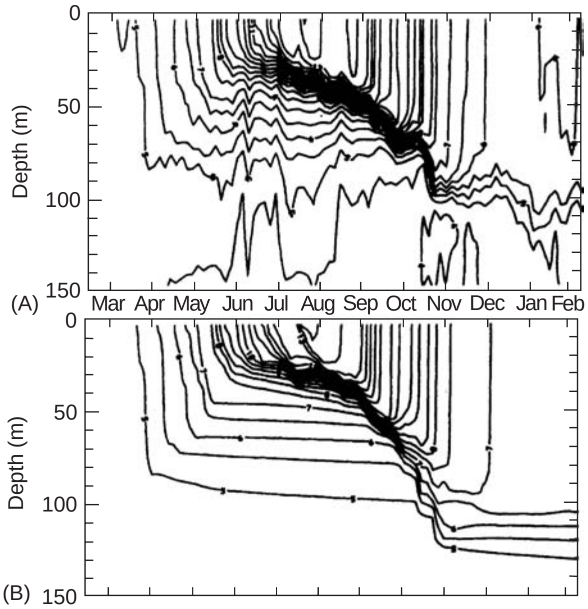

Figure 8 Comparison of (A) observed and (B) one-dimensional model-simulated mixed layer temperatures at Ocean Station PAPA in the North-Eastern Pacific. Simulations are remarkably good, given the large uncertainties associated with forcing parameters and the neglect of advection effects. (Kantha and Clayson, 2000.)

partial differential equations governing second moments in addition to the usual ones for mass, momentum (for velocity components U and V), and scalar (for temperature T and salinity S) conservation. Attempts have therefore been made to simplify the set by once again utilizing certain aspects of turbulence such as its departure from the state of local isotropy. The result is a hierarchy of models, of which the most useful for geophysical applications is the model that consists of one conservation equation for TKE (half the square of the turbulence velocity macroscale q), and a set of algebraic equations for turbulence second-moment quantities. The resulting simplicity and the potential 'universality' of application are of particular interest in modeling OMLs. Since the most basic description

of turbulence is incomplete without quantifying its macro length scale l, this set is supplemented either by auxiliary information on the turbulence length scale or else by utilizing an equation for a quantity that includes the turbulence length scale such as dissipation (q^3/l) or the product of the turbulence length scale and twice the TKE (q^2l). **Figure 8** shows an example of the accuracy attainable with a current-generation OML model.

To summarize, there has been a significant increase in our understanding of the OML in the last two decades. However, much work remains. A better understanding and quantification of the gravity-wave- and buoyancy- and momentum-flux-driven near-surface layers still remain elusive.

See also

Air–Sea Interaction: Surface Waves. **Boundary Layers:** Convective Boundary Layer; Neutrally Stratified Boundary Layer; Stably Stratified Boundary Layer. **Ocean Circulation:** General Processes; Water Types and Water Masses. **Parameterization of Physical Processes:** Turbulence and Mixing.

Further Reading

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Stably Stratified Boundary Layer

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Introduction

Stable boundary layers are most commonly generated by surface radiative cooling or advection of warm air over a cooler surface. Turbulence transports heat toward the cooler surface. The actual transfer of heat to the surface occurs by thermal conductivity through a thin laminar sublayer, perhaps only a few millimeters thick. However, the rate of this heat conduction is dictated by the downward turbulent transport of heat to the laminar sublayer. Therefore, we concentrate on the turbulent transfer.

The downward transport of heat corresponds to downward buoyancy flux that destroys turbulent kinetic energy. That is, the energy required to push