Okhotsk Sea products, then enter the interior of the North Pacific.

## See also

**Bottom Water Formation. Kuroshio and Oyashio Currents. Polynyas. Sea Ice: Overview; Variations in Extent and Thickness. Thermohaline Circulation. Tides. Tidal Energy. Upper Ocean Mixing Processes. Water Types and Water Masses. Wind Driven Circulation.**

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# OPEN OCEAN CONVECTION

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## Introduction

Free convection is fluid motion due to buoyancy forces. Free convection, also referred to as simply convection, is driven by the static instability that results when relatively dense fluid lies above relatively light fluid. In the ocean, greater density is associated with colder or saltier water, and it is possible to have thermal convection due to the vertical temperature gradient, haline convection due to the vertical salinity gradient, or thermohaline convection due to the combination.

Since sea water is about 1000 times denser than air, the air–sea interface from the waterside can be considered a free surface. So-called thermocapillary convection can develop near this surface owing to the dependence of the surface tension coefficient on temperature. There are experimental indications that in the upper ocean layer more than 2 cm deep, buoyant convection dominates. Surfactants, however, may affect in the surface renewal process. This article will mainly consider convection without these capillary effects.

Over most of the ocean, the near-surface region is considered to be a mixed layer in which turbulent mixing is stronger than at greater depth. The strong mixing causes the mixed layer to have very small vertical variations in density, temperature, and other properties compared to the pycnocline region below. Convection is one of the key processes driving mixed layer turbulence, though mechanical stirring driven by wind stress and other processes is also important. Therefore, understanding convection is crucial to understanding the mixed layer as well as property fluxes between the ocean and the atmosphere.

Thermal convection is associated with the cooling of the ocean surface due to sensible ($Q_T$), latent ($Q_L$), and effective long-wave radiation ($Q_E$) heat fluxes. $Q_T$ may have either sign; its magnitude is, however, much less than that of $Q_E$ or $Q_L$ (except perhaps in some extreme situations). The top of the water column becomes colder and denser than the water below, and convection begins. In this way, cooling is associated with the homogenization of the water column and the deepening of the mixed layer. Warming due to solar radiation occurs in the surface layer of the ocean and is associated with restratification and reductions in mixed layer depth. The most prominent examples of this mixing/restratification process are the diurnal cycle (night-time cooling and daytime warming) and the seasonal cycle (winter cooling and summer warming).

There are also important geographical variations in convection, with net cooling of relatively warm water occurring more at higher latitudes and a net warming of water occurring closer to the Equator. For this reason, mixed layer depth generally increases towards the poles, though at very high latitudes ice-melt can lower the surface salinity enough to inhibit convection. Over most of the ocean, annual average mixed layer depths are in the range of 30–100 m, though very dramatic convection in such places as the Labrador Sea, Greenland Sea, and western Mediterranean Sea can deepen the mixed layer to thousands of meters. This article discusses convection reaching no deeper than a few hundred meters. Dynamically, the convection discussed here differs from deep convection in being more strongly affected by surface wind stress and much less affected by the rotation of the Earth.

Convection directly affects several aspects of the near-surface ocean. Most obviously, the velocity patterns of the turbulent flow are influenced by the presence of convection, as is the velocity scale. The convective velocity field then controls the vertical transport of heat (or more correctly, internal energy), salinity, momentum, dissolved gases, and other properties, and the vertical gradients of these properties within the mixed layer. Convection helps to determine property exchanges between the atmosphere and ocean and the upper ocean and the deep ocean. The importance of convection for heat and gas exchange has implications for climate studies, while convective influence on the biologically productive euphotic zone has biological implications as well.

## Phenomenology

The classical problem of free convection is to determine the motion in a layer of fluid in which the top surface is kept colder than the bottom surface. This is an idealization of such geophysical examples as an ocean being cooled from above or the atmosphere being heated from below. The classical problem ignores such complications as wind stress on the surface, waves, topographic irregularities, and the presence of a stably stratified region below the convection region. The study of convection started in the early twentieth century with the experiments of Benard and the theoretical analysis of Rayleigh. One might expect that heavier fluid would necessarily exchange places with lighter fluid below as a result of buoyancy forces. This happens by means of convective cells or localized plumes of sinking dense fluid and rising light fluid. However, such cells or plumes are retarded by viscous forces and are also dissipated by thermal diffusion as they sink into an environment with a different density. When the buoyancy force is not strong enough to overcome the inhibitory effects, the heavy-over-light configuration is stable and no convection forms. The relative strengths of these conflicting forces is measured by the Rayleigh number, a nondimensional number given by eqn [1].

$$Ra = \frac{g\alpha\Delta T h^3}{(k_T \nu)} \qquad [1]$$

Here $g$ is the acceleration of gravity, $\alpha$ is the thermal expansion coefficient of sea water ($\alpha = 2.6 \times 10^{-4}\,^\circ\mathrm{C}^{-1}$ at $T = 20^\circ\mathrm{C}$ and $S = 35\,\mathrm{PSU}$), $\Delta T$ is the temperature difference between the top and bottom surfaces, $h$ is the convective layer thickness, and $\nu$ and $k_T$ are the molecular coefficients of viscosity and thermal diffusivity, respectively ($\nu = 1.1 \times 10^{-6}\,\mathrm{m^2\,s^{-1}}$ and $k_T = 1.3 \times 10^{-7}\,\mathrm{m^2\,s^{-1}}$ at $T = 20^\circ\mathrm{C}$ and $S = 35\,\mathrm{PSU}$). The term $\alpha\Delta T = \Delta\rho/\rho$ represents the fractional density difference between top and bottom.

Convection occurs only if $Ra$ is greater than a critical value, $Ra_{cr}$, which depends somewhat on geometrical and other details of the fluid. For the classical problem of water bounded above and below by solid plates, $Ra_{cr} = 657$. For sea water under typical conditions, even a temperature difference of $0.1^\circ\mathrm{C}$ makes $Ra > Ra_{cr}$ as long as $h$ is greater than a centimeter.

For $Ra > Ra_{cr}$, the Rayleigh number still serves a useful purpose as a guide to the nature of the convective activity (though the problem also depends on the Prandtl number, $Pr = \nu/k_T$). For

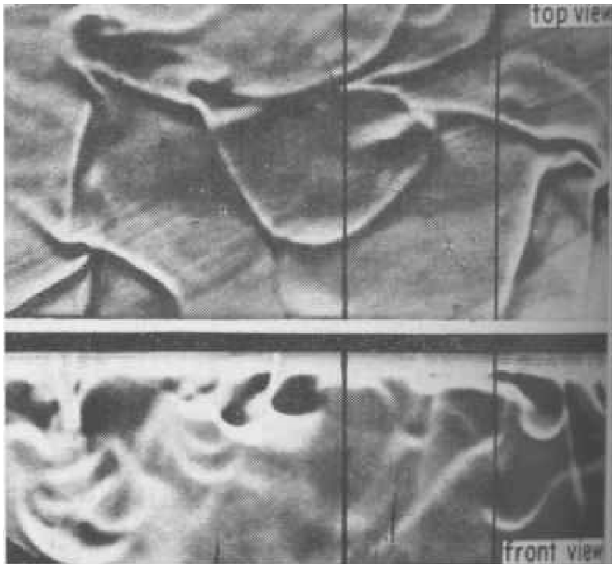


**Figure 1** Orthogonal views of convective streamers in warm water that is cooling from the surface. The constantly changing patterns appear as intertwining streamers in the side view. (From Spangenberg WG and Rowland WR (1961)) Convective circulation in water induced by evaporative cooling. *Physics of Fluids* 4: 743–750. © 1961 American Institute of Physcis.

a fixed *Pr* and for *Ra* only slightly larger than $Ra_{cr}$, motion occurs in regular, steady cells. As *Ra* is increased, the motion becomes time-dependent. Regular oscillations occur, and these increase in number and frequency for higher *Ra*. At sufficiently high *Ra*, the flow is turbulent and intermittent. The value of *Ra* in the ocean is very large (typically greater than $10^{14}$ for a temperature difference of 0.1°C over 10 m), so convection is usually turbulent.

Turbulent convection is usually characterized by the formation of descending parcels of cold water. In laboratory experiments, it has been found that water from the cooled surface layer collects along lines, producing thickened regions that become unstable and plunge in vertical sheets (**Figure 1**). In analogy to atmospheric convection, we will here call these parcels thermals, although — in contrast to the atmosphere — in the ocean they are colder than the surrounding fluid. In 1966, Howard formulated a phenomenological theory that represented turbulent convection as the following cyclic process. The thermal boundary layer forms by diffusion, grows until it is thick enough to start convecting, and is destroyed by convection, which in turn dies down once the boundary layer is destroyed. Then the cycle begins again. This phenomenological theory has implications for the development of parametrizations for the air–sea heat and gas exchange under low wind speed conditions (see later).

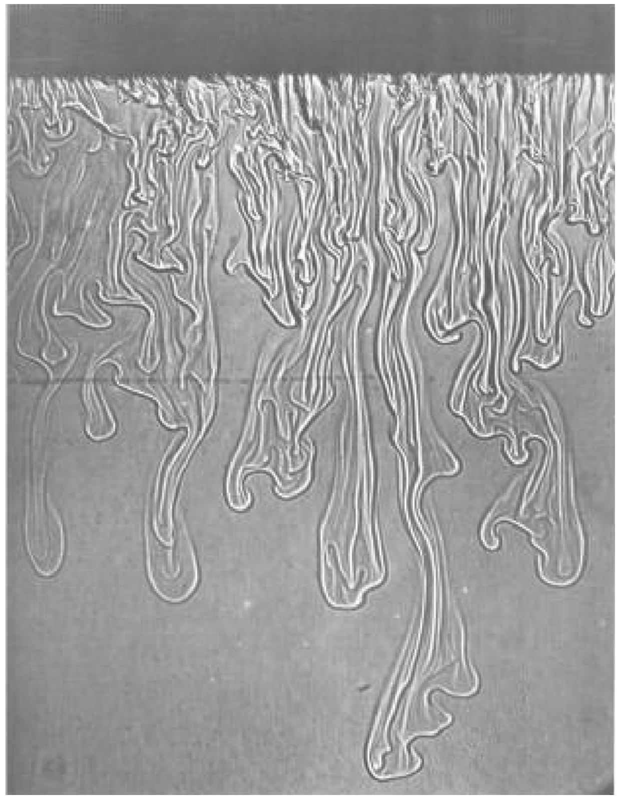

**Figure 2** Shadowgraph picture of the development of secondary haline convection. From Foster TD (1974) The hierarchy of convection. *Colloques Internationaux du CNRS N215.* Processus de Formation Des Eaux Oceaniques Profondes, pp. 235–241. © 1974 Centre National de la Recherche Scientifique.

The descending parcels of water have a mushroomlike appearance. In the process of descending to deeper layers, the descending parcels developing as a result of the local convective instability of the thermal molecular sublayer join and form larger mushroomlike structures. The latter descend faster and eventually form bigger structures. This cascade process produces a hierarchy of convective scales, which is illustrated in **Figure 2** on the example of the haline convection.

## Penetrative Convection

The unstable stratification of the mixed layer is usually bounded below by a stratified pycnocline. One can imagine the mixed layer growing in depth with thermals confined to the statically unstable depth range. Suppose the density at the top of the pycnocline is $\rho_1$ (**Figure 3A**). As surface buoyancy loss and convection increase the average density of the mixed layer, the mixed layer density increases to $\rho_2$, which is slightly denser than $\rho_1$ (**Figure 3B**). The static instability now allows convection to act on the

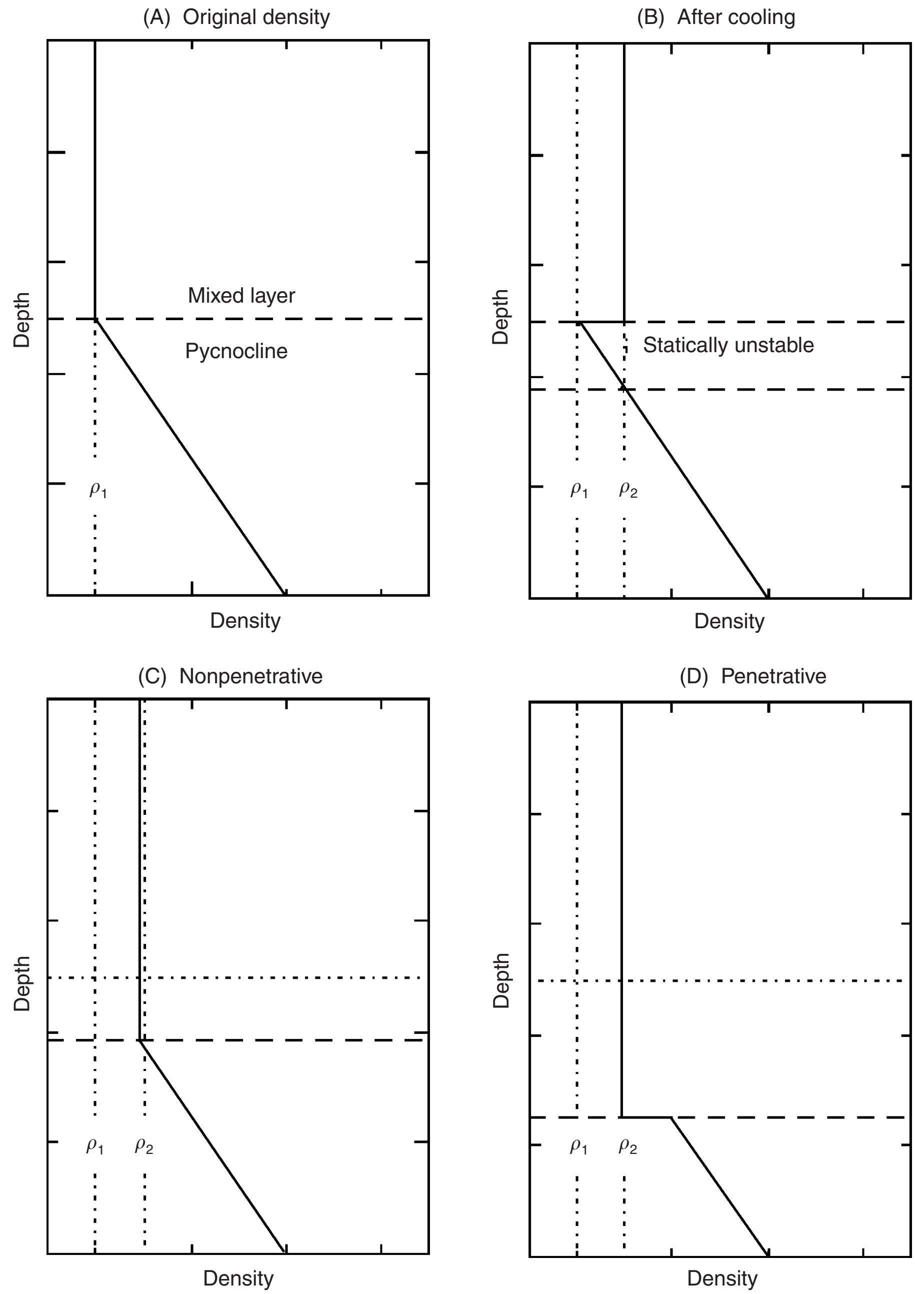


**Figure 3** Schematic diagram of nonpenetrative and penetrative convection.

pycnocline down to density $\rho_2$ (**Figure 3C**), so that the mixed layer grows at the expense of the pycnocline. This is known as nonpenetrative convection.

In reality, the largest thermals acquire enough kinetic energy, as they fall through the mixed layer, that they can overshoot the base of the mixed layer, working against gravity. This is penetrative convection. The penetrative convection produces a countergradient flux that is not properly accounted for if we model convective mixing as merely a very strong vertical diffusion. Unlike the smooth density profile at the base of a mixed layer that is growing by nonpenetrative convection (**Figure 3C**), penetrative convection is characterized by a density jump at the base of the mixed layer (**Figure 3D**).

The cooling of the ocean from its surface is compensated by the absorption of solar radiation. The latter is a volume source for the upper meters of the ocean. The thermals from the ocean surface, as they descend deeper into the mixed layer, produce heat flux that is compensated by the volume absorption of solar radiation. This is another type of the penetrative convection in the upper ocean, which will be considered in more detail in a later section.

## Relative Contributions of Convection and Shear Stress to Turbulence

For the limiting case in which the only motion in the mixed layer is due to convection, there are simple estimates of average speed and temperature fluctuations associated with the plumes. When the Rayleigh number is high enough that the flow is fully turbulent, the plume characteristics should be largely independent of the viscosity and diffusivity throughout most of the mixed layer. In that case, ignoring the Earth's rotation and influences from the pycnocline, the governing parameters of the system are simply the mixed layer depth $h$ and the surface buoyancy flux $B_0$. $B_0$ is based on the surface heat fluxes according to eqn [2], where $\rho$ is the water density, $c_p$ is the specific heat capacity of water ($\approx 4\times10^3\,\mathrm{J\,kg^{-1}\,K^{-1}}$), $L$ is the latent heat released by evaporation ($\approx 2.5\times10^6\,\mathrm{J\,kg^{-1}}$), $S$ is the surface salinity, and $\beta$ is the coefficient of salinity expansion ($\beta = 7.4\times10^{-4}\,\mathrm{PSU^{-1}}$ at $T = 20^\circ\mathrm{C}$ and $S = 35\,\mathrm{PSU}$).

$$B_0 = -g\rho^{-1}[\alpha c_p^{-1}(Q_\mathrm{S} + Q_\mathrm{E} + Q_\mathrm{L}) + \beta Q_\mathrm{L} L^{-1} S] \quad [2]$$

The first term in the square bracket in the right side of eqn [2] relates to the buoyancy flux due to surface cooling; the second term relates to the buoyancy flux due to the surface salinity increase because of evaporation.

Given all the above restrictions, the velocity scale, $w_*$, is then given by the Priestly formula (eqn [3]).

$$w_* = (B_0 h)^{1/3} \quad [3]$$

This is the only combination of $B_0$ and $h$ that will give the proper units for velocity. Similarly, if we define the buoyancy to be $b = g\Delta\rho/\rho$, the buoyancy scale, $b_*$, is given in eqn [4].

$$b_* = (B^2/h)^{1/3} \quad [4]$$

Laboratory experiments have shown that these scales are in good agreement with actual fluctuations during convection. For typical oceanic parameters (for instance, heat flux of $Q_0 = 100\,\mathrm{W\,m^{-2}}$ and $h = 100\,\mathrm{m}$), $w_*$ is a few centimeters per second and $b_*$ is equivalent to temperature fluctuations of about $0.01^\circ\mathrm{C}$.

Two major sources of turbulent kinetic energy in the upper ocean are the wind stress and buoyant convection. Upper ocean convection is usually accompanied by near-surface currents induced by wind and wind waves. The near-surface shear is then an additional source of near-surface turbulent mixing. In the 1950s, Oboukhov proposed the buoyancy length scale, $L_\mathrm{O} = \kappa u_*^3/B_0$, where $\kappa$ is the Von Karman constant ($\kappa = 0.4$), $B_0$ is the surface buoyancy flux (e.g., defined by [2]), and $u_*$ is the boundary layer velocity scale (friction velocity) defined as $u_* = (\tau/\rho)^{1/2}$, where $\tau$ represents the bottom stress in the atmospheric case and the windstress in the oceanic case ($\rho$ is the density of air or water, respectively). Later, Monin and Oboukhov suggested the stability parameter, $\xi = z/L_\mathrm{O}$ (where $z$ is the height in the atmosphere or the depth in the ocean), to characterize the relative importance of shear and buoyant convection in the planetary boundary layer. Experimental studies conducted in the atmospheric boundary layer show that at $\xi < -0.1$ the flow is primarily driven by buoyant convection. From the analogy between the atmospheric and oceanic turbulent boundary layers, the Monin–Oboukhov theory is often applicable to the analysis of the oceanic processes as well. In particular, it provides us a theoretical basis to separate the layers of free and forced convection in the upper ocean turbulent boundary layer.

For a $5\,\mathrm{m\,s^{-1}}$ wind speed and $Q_0 = 100\,\mathrm{W\,m^{-2}}$, the Oboukhov scale is $L_\mathrm{O} \sim 15\,\mathrm{m}$. This means that the shear-driven turbulent flow is confined within a 1.5 m thick near-surface layer of the ocean. In a 50 m deep mixed layer, 97% of its depth will be driven by the buoyant convection during nighttime, with the rate of dissipation of turbulent kinetic energy there about equal to the surface buoyancy flux, $B_0$, as shown by Shay and Gregg.

## Convection and Molecular Sublayers

Convection is driven by the horizontal–mean vertical density gradient. At high $Ra$, typical vertical velocities are much lower near the top and bottom boundaries than they are in the bulk of the water column. Since the vertical density gradient is reduced by the convective motion, the velocity distribution causes most of the vertical density gradient to occur near the boundaries. Indeed, under low-wind, low-wave conditions in which convection dominates, the mixed layer temperature gradient is largely confined to a region only about 1 mm deep. Because the vertical heat flux at the base of the convection region is typically much smaller than at the surface, the large temperature gradient only occurs at the surface, where this thermal sublayer is often referred to as the cool skin.

The temperature jump across the cool skin can be related to the vertical flux of heat at the air–sea interface and constants of molecular viscosity and heat diffusion in water using convection laws. The vertical heat flux, $Q_0$, can be written in nondimensional form as the Nusselt number (eqn [5]).

$$Nu = (Q_0/c_p\rho)/(k_\mathrm{T}\Delta T/h) \qquad [5]$$

In eqn [5] the heat flux is normalized by the heat flux due to vertical diffusion in the absence of convection. This quantity must be a function of the given nondimensional parameters of the system, which, for thermal convection in the absence of other driving mechanisms, are just $Ra$ and the Prandtl number $Pr$ (here we ignore the Earth's rotation and entrainment from the pycnocline). A further simplifying assumption is that for high $Ra$ (greater than $10^7$), typical of the mixed layer, the convection is fully turbulent and does not depend on the mixed layer thickness, $h$, which implies eqn [6], where $A(Pr)$ is a dimensionless coefficient depending on Prandtl-number (according to laboratory measurements, $A \approx 0.16$–$0.25$).

$$Nu = A(Pr)Ra^{1/3} \qquad [6]$$

Given the definitions of $Ra$ and $Nu$, this relation can be rearranged to yield the temperature difference across the cool skin, $\Delta T$, as a function of the surface heat flux, $Q_0 = Q_\mathrm{L} + Q_\mathrm{E} + Q_\mathrm{T}$ (eqn [7]).

$$\Delta T = A^{-3/4}(\alpha g k_\mathrm{T}^2/\nu)^{-1/4}(Q_0/c_p\rho)^{3/4} \qquad [7]$$

$\Delta T$ is 0.2–0.4°C under typical oceanic conditions but can be as much as 1°C in regions of very high heat loss to the atmosphere (e.g., Gulf Stream at high latitudes).

While the term 'sea surface temperature' (SST) is often used to represent the temperature of the mixed layer as a whole, the existence of a cool skin means that the temperature of the literal surface of the ocean can be somewhat lower than the rest of the mixed layer. Satellite measurements of SST are based on infrared emissions from a thin layer of several micrometers, so that these measurements can be somewhat different from ship-based 'surface' measurements, which are generally based on sampling within several upper meters of the ocean. Indeed, while the first experimental evidence of the cool skin was obtained in the 1920s, the phenomenon was not widely recognized by the oceanic community until sophisticated methods, including remote sensing by infrared techniques, had been gradually helping to incorporate the cool skin into modern oceanography.

The accuracy of current satellite remote sensing techniques is, nevertheless, still below that level at which the cool skin becomes of crucial importance. The effect of the cool skin on the heat exchange between ocean and atmosphere is also basically below the resolution of widely used bulk flux algorithms. However, one interesting practical application of the cool skin phenomenon emerged in the 1990s. Similar laws govern the thermal sublayer of the ocean (the cool skin) and diffusive sublayers associated with air–sea gas exchange. Such gas exchange is a key biogeochemical variable, and for greenhouse gases such as $CO_2$ is of climatological importance as well. The rate at which gases cross the air–sea interface is measured by the piston velocity, $K$. Boundary layer laws relate $K$ to $\Delta T$ in the eqn [8].

$$K = \frac{A_0 Q_0(\mu/\nu)^{1/2}}{(c_p\rho\Delta T)} \qquad [8]$$

$A_0$ is a dimensionless constant ($\approx 1.85$) and $\mu$ is the molecular gas diffusion coefficient in water ($\mu = 1.6\times10^{-9}\,\mathrm{m^2\,s^{-1}}$ for $CO_2$ at $T = 20$°C and $S = 35$ PSU). The more readily available cool skin data can then be used for an adjustment of the gas transfer parametrization.

The convective parametrizations for the cool skin and air–sea gas exchange are valid within the range of wind speed from 0 to 3–4 $\mathrm{m\,s^{-1}}$. Under higher wind speed conditions, the cool skin and the interfacial air–sea gas exchange are controlled by the wind stress and surface waves. The transition is observed when the surface Richardson number, $Rf_0^i = \alpha g Q_0/(c_p\rho u_*^4)$, reaches a value of approximately $-1.5\times10^{-5}$ (here $\rho$ is the water density).

## Diurnal and Seasonal Cycles of Convection

For much of the year, much of the ocean experiences a cycle of daytime heating and nighttime cooling that leads to a strong diurnal cycle in convection and mixed layer depth. Such behavior is illustrated in **Figure 4**. At night, when there is cooling, the convective plumes reach the base of the mixed layer, which deepens as the mixed layer grows colder and denser. During the day, convection is inhibited within the bulk of the mixed layer but may still occur near the surface of the mixed layer, even if the mixed layer experiences a net heat gain. This is

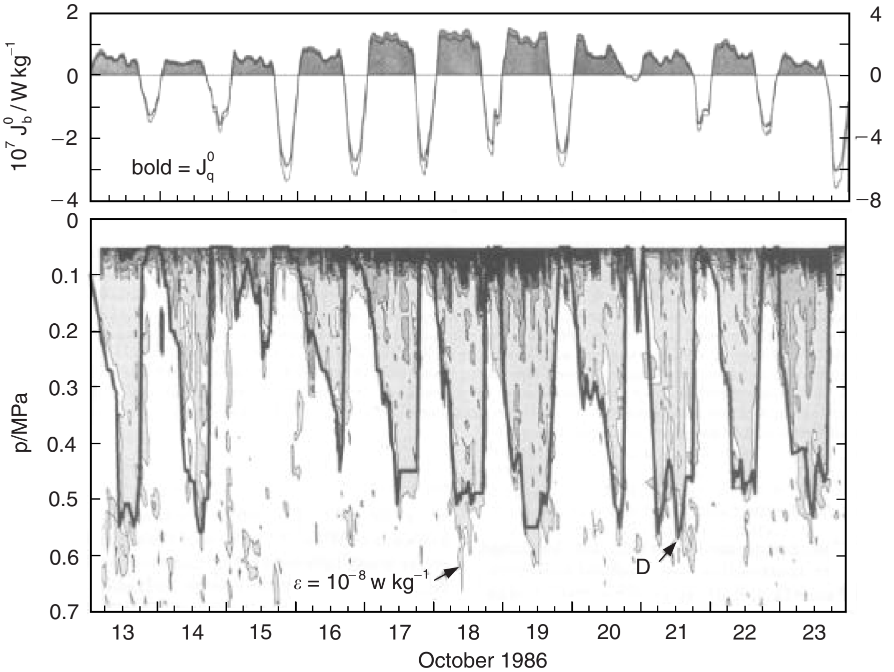


**Figure 4** Diurnal cycles in the outer reaches of the California Current (34°N, 127°W). Each day the ocean lost heat and buoyancy starting several hours before sunset and continuing until a few hours after sunrise. These losses are shown by the shaded portions of the surface heat and buoyancy fluxes in the top panel. In response, the surface turbulent boundary layer slowly deepened (lower panel). The solid line marks *D*, the depth of the surface turbulent boundary layer, and the lightest shading shows $10^{-8}\,\mathrm{W\,kg^{-1}} < \varepsilon < 10^{-7}\,\mathrm{W\,kg^{-1}}$ where $\varepsilon$ is the dissipation rate of the turbulent kinetic energy. The shading increases by decades, so that the darkest shade is $\varepsilon > 10^{-5}\,\mathrm{W\,kg^{-1}}$. Note that 1 MPa in pressure *p* corresponds to approximately 100 m in depth, $J_b^0 = -B_0$, and $J_q^0 = -(Q_0 + Q_R)$, where $Q_R$ is the solar radiation flux penetrating ocean surface. (From Lombardo CP and Gregg MC (1989) Similarity scaling during nighttime convection. *Journal of Geophysical Research* 94: 6273–6274.) © 1989 American Geophysical Union.

because the vertical distribution of cooling and heating are somewhat different. Heat loss is dominated by latent heat flux associated with evaporation and hence this forcing occurs at the top surface. Heat gain is dominated by solar radiation that is absorbed by the water over a range of depths that can extend tens of meters in many parts of ocean. For example, one can have surface heat loss of $100\,\mathrm{W\,m^{-2}}$ occurring at the surface and net radiative heat gain of $500\,\mathrm{W\,m^{-2}}$ distributed over the top 30 m of the ocean. Calculating the rate of change of heat due to the forcing between the surface and a depth $z$, we find that there is actually heat loss for small $z$ down to a depth known as the thermal compensation depth. Below this depth, the mixed layer restratifies and convection occurs only through the mechanism of penetrative convection. For most of the World Ocean, the thermal compensation depth is less than 1 m between sunrise and sunset.

Usually, the rate of turbulent kinetic energy production in the mixed layer is dominated by the convective term at night, but by the wind stress term during most of the day. Because the thermal compensation depth is generally quite small, turbulent kinetic energy generated by convection makes no contribution to turbulent entrainment of water through the bottom of the mixed layer, which lies much deeper. Under low wind speed conditions and strong solar insolation, the thickness of the surface convective layer of the ocean may reduce to only several centimeters. In that case, convection in the upper ocean may be of a laminar or transitional nature.

Stable stratification inhibits turbulent mixing below the relatively thin near-surface convection layer. Vertical mixing of momentum is confined to the shallow daytime mixed layer so that, during the day, flow driven directly by the wind stress is confined to a similarly thin current known as the diurnal jet. In the evening, when convection is no longer confined by the solar radiation effect, convective plumes penetrate deeper into the stratified part of the mixed layer, increasing the turbulent mixing of momentum at the bottom of the diurnal jet. The diurnal jet then releases its kinetic energy during a relatively short time. This process is so intensive that the releasing kinetic energy cannot be dissipated locally. As a result, a Kelvin–Helmholtz type instability is formed, which generates billows — a kind of organized structure. The billows intensify the deepening of the diurnal mixed layer.

Although the energy of convective elements is relatively small, it serves as a catalyst for the release of the kinetic energy by the mean flow. In the equatorial ocean, the shear in the upper ocean is intensified by the Equatorial Undercurrent; the evening deepening of the diurnal jet is therefore sometimes so intense that it resembles a shock, which radiates very intense high-frequency internal waves in the underlying thermocline.

The diurnal cycle is often omitted from numerical ocean models for reasons of computational cost. However, the mixed layer response to daily-averaged surface fluxes is not necessarily the same as the average response to the diurnal cycle. Neglecting the diurnal cycle replaces periodic nightly convective pulses with chronic mixing that does not reach as deep.

Open ocean convection is a mechanism effectively controlling the seasonal cycle in the ocean as well. Resolution of diurnal changes is usually uneconomical when the seasonal cycle is considered. Because of nonlinear response of the upper ocean to the atmospheric forcing, simply averaged heat fluxes cannot be used to estimate the contribution of the convection on the seasonal scale. The sharp transition between the nocturnal period, when convection dominates mixing in the surface layer, and the daytime period, when the sun severely limits the depth of convection, leaving the wind stress to control mixing, may simplify the design of models for the seasonal cycle of the upper ocean. Incorporation of convection adjustment schemes into the oceanic component of the global circulation models leads to an appreciable change of troposphere temperature in high latitudes, which affects the global ocean and atmosphere circulation. Parametrization of the convection on the seasonal and global scales is therefore an important task for the prediction of climate and its changes.

## Conclusions

Observation of the open ocean convection is a difficult experimental task. Though convective processes have been observed in several oceanic turbulence studies, most of our knowledge of this phenomenon in the ocean is based on the analogy between atmospheric and oceanic boundary layers and on laboratory studies. Many intriguing questions regarding the convection in the open ocean remain, however. Some of them, like the role of penetrative convection in mixed layer dynamics, are of crucial importance for improvement of the global ocean circulation modeling. Others, like the role of surfactants in the surface renewal process, are of substantial interest for studying the air–sea exchange and global balance of greenhouse gases like $CO_2$.

## See also

**Air–Sea Gas Exchange. Breaking Waves and Near-surface Turbulence. Deep Convection. Non-rotating Gravity Currents. Photochemical Processes. Rotating Gravity Currents. Satellite Remote Sensing of Sea Surface Temperatures. Small-scale Patchiness, Models of. Single Compound Radiocarbon Measurements. Three-dimensional (3D) Turbulence. Upper Ocean Mixing Processes. Wind and Buoyancy-forced Upper Ocean.**

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