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

See also

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

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UPPER OCEAN MIXING PROCESSES

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Introduction

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

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

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

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

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

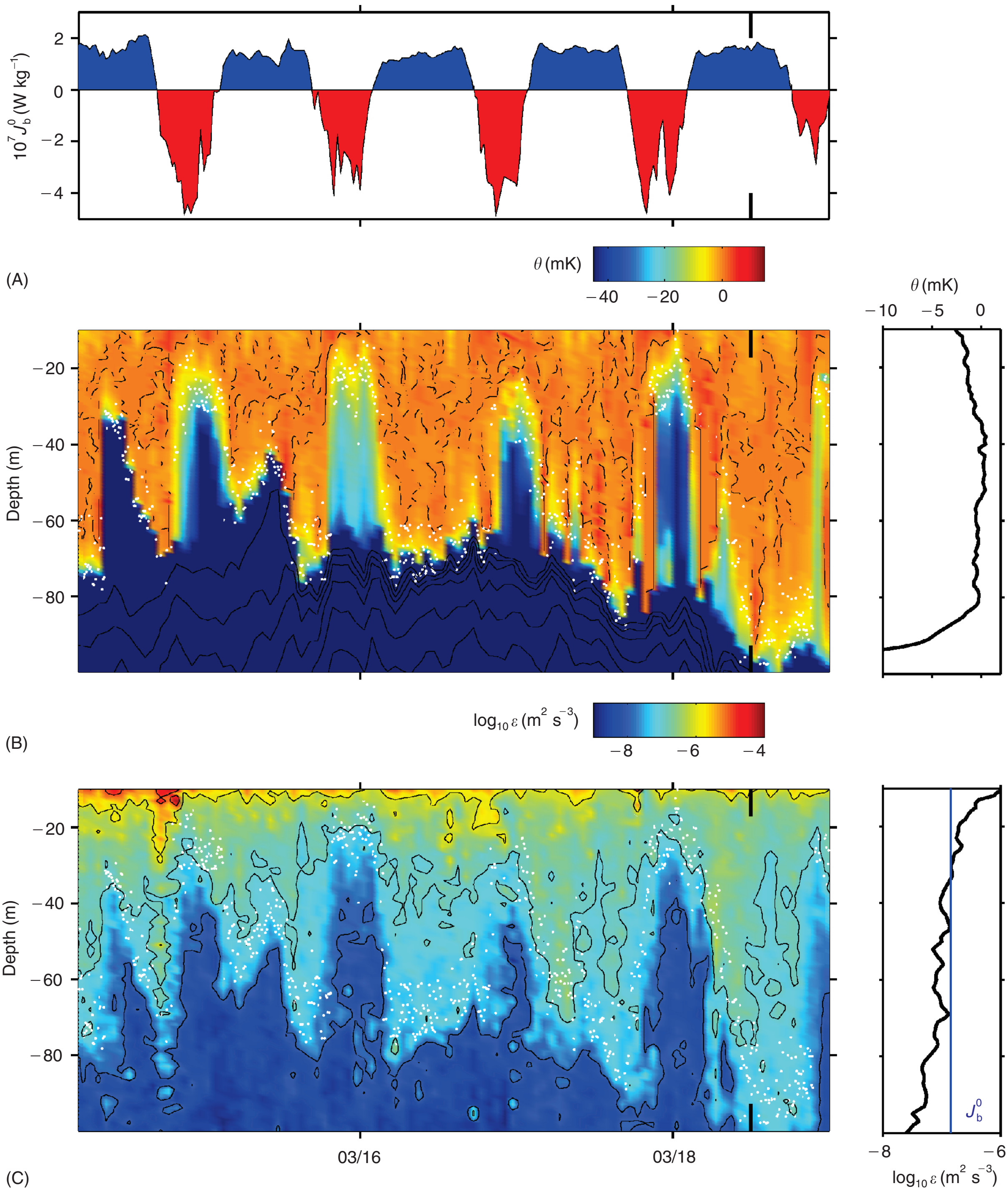

Figure 1 Observations of mixing in the upper ocean over a five-day period. These observations were made in March 1987 in the North Pacific using a vertical turbulence profiler and shipboard meteorological sensors. (A) The variation in the surface buoyancy flux, J_b^0, which is dominated by surface heating and cooling. The red (blue) areas represent daytime heating (nighttime cooling). Variations in the intensity of nighttime cooling are primarily due to variations in winds. (B) Potential temperature referenced to the individual profile mean in order to emphasize vertical rather than horizontal structure (θ; K). To the right is an averaged vertical profile from the time period indicated by the vertical bars at top and bottom of each of the left-hand panels. (C) The intensity of turbulence as indicated by the viscous dissipation rate of turbulence kinetic energy, ε. To the right is an averaged profile with the mean value of J_b^0 indicated by the vertical blue line. The dots in (B) and (C) represent the depth of the mixed layer as determined from individual profiles.

Below this superadiabatic surface layer is a layer of depth 10–30 m in which the temperature change is less than 1 mK. Within this mixed layer, the intensity of turbulence, as quantified by the turbulent kinetic energy dissipation rate, ε, is relatively uniform and approximately equal to J_b^0. (ε represents the rate at which turbulent motions in a fluid are dissipated to heat. It is an important term in the evolution equation for turbulent kinetic energy, signifying the tendency for turbulence to decay in the absence of forcing.) Below the mixed layer, ε generally (but not always) decreases, whereas above, ε increases by 1–2 factors of 10.

Below the mixed layer is a region of stable stratification that partially insulates the upper ocean from the ocean interior. Heat, momentum, and chemical species exchanged between the atmosphere and the ocean interior must traverse the centimeters thick cool skin at the very surface, the surface layer, and the mixed layer to modify the stable layer below. These vertical transports are governed by a combination of processes, including those that affect only the surface itself (rainfall, breaking surface gravity waves), those that communicate directly from the surface throughout the entire mixed layer (convective plumes) or a good portion of it (Langmuir circulations) and also those processes that are forced at the surface but have effects concentrated at the mixed-layer base (inertial shear, Kelvin–Helmholtz instability, propagating internal gravity waves). Several of these processes are represented in schematic form in **Figure 2**. Whereas **Figure 1** represents the observed time evolution of the upper ocean at a single location, **Figure 2** represents an idealized three-dimensional snapshot of some of the processes that contribute to this time evolution.

Heating of the ocean's surface, primarily by solar (short-wave) radiation, acts to stabilize the water column, thereby reducing upper ocean mixing. Solar radiation, which peaks at noon and is zero at night, penetrates the air–sea interface (limited by absorption and scattering to a few tens of meters), but heat is lost at the surface by long-wave radiation, evaporative cooling and conduction throughout both day and night. The ability of the atmosphere to modify the upper ocean is limited by the rate at which heat

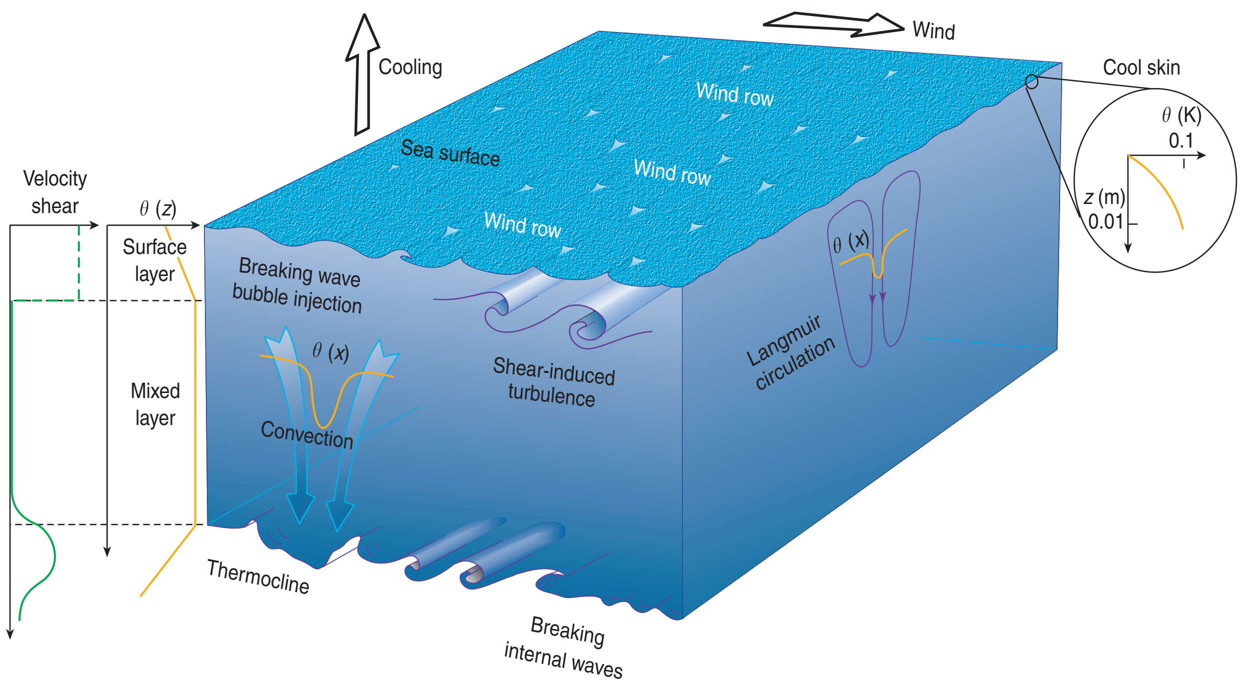

Figure 2 Diagram showing processes that have been identified by a wide range of observational techniques as important contributors to mixing the upper ocean in association with surface cooling and winds. The temperature (θ) profiles shown here have the adiabatic temperature (that due to compression of fluid parcels with depth) removed; this is termed potential temperature. The profile of velocity shear (vertical gradient of horizontal velocity) indicates no shear in the mixed layer and nonzero shear above. The form of the shear in the surface layer is a current area of research. Shear-induced turbulence near the surface may be responsible for temperature ramps observed from highly resolved horizontal measurements. Convective plumes and Langmuir circulations both act to redistribute fluid parcels vertically; during convection, they tend to move cool fluid downward. Wind-driven shear concentrated at the mixed-layer base (thermocline) may be sufficient to allow instabilities to grow, from which internal gravity waves propagate and turbulence is generated. At the surface, breaking waves inject bubbles and highly energetic turbulence beneath the sea surface and disrupt the ocean's cool skin, clearing a pathway for more rapid heat transfer into the ocean.

and momentum can be transported across the air–sea interface. The limiting factor here is the viscous boundary layer at the surface, which permits only molecular diffusion through to the upper ocean. This layer is evidenced by the ocean's cool skin, a thin thermal boundary layer (a few millimeters thick), across which a temperature difference of typically 0.1 K is maintained. Disruption of the cool skin permits direct transport by turbulent processes across the air–sea interface. Once disrupted, the cool skin reforms over a period of some tens of seconds. A clear understanding of processes that disrupt the cool skin is crucial to understanding how the upper ocean is mixed.

Convection

Cooling at the sea surface creates parcels of cool, dense fluid, which later sink to a depth determined by the local stratification in a process known as convection. Cooling occurs almost every night and also sometimes in daytime in association with weather systems such as cold air outbreaks from continental landmasses. Convection may also be caused by an excess of evaporation over precipitation, which increases salinity, and hence density, at the surface. Winds aid convection by a variety of mechanisms that agitate the sea surface, thereby disrupting the viscous sublayer and permitting rapid transfer of heat through the surface (see below). Convection in the ocean is analogous to that found in the daytime atmospheric boundary layer, which is heated from below, and which has been studied in great detail. Recourse to atmospheric studies of convection has helped in understanding the ocean's behavior.

Surface tension and viscous forces initially prevent dense, surface fluid parcels from sinking. Once the fluid becomes sufficiently dense, however, these forces are overcome and fluid parcels sink in the form of convective plumes. The relative motion of the plumes helps to generate small-scale turbulence, resulting in a turbulent field encompassing a range of scales from the depth of the mixed layer (typically 100 m) to a few millimeters.

A clear feature of convection created by surface cooling is the temperature profile of the upper ocean (**Figure 1**). Below the cool skin is an unstable surface layer that is the signature of plume formation. Below that is a well-mixed layer in which density (as well as temperature and salinity) is relatively uniform. The depth of convection is limited by the local thermocline. Mixing due to penetrative convection into the thermocline represents another source of cooling of the mixed layer above. Within the convecting layer, there is an approximate balance between buoyant production of turbulent kinetic energy and viscous dissipation, as demonstrated by the observation $\varepsilon \approx J_b^0$.

The means by which the mixed layer is restratified following nighttime convection are not clear. Whereas some one-dimensional models yield realistic time series of sea surface temperature, suggesting that restratification is a one-dimensional process (see below), other studies of this issue have shown one-dimensional processes to account for only 60% of the stratification gained during the day. It has been suggested that lateral variations in temperature, due to lateral variations in surface fluxes, or perhaps lateral variations in salinity due to rainfall variability, may be converted by buoyant forces into vertical stratification. These indicate the potential importance of three-dimensional processes to restratification.

Wind Forcing

Convection is aided by wind forcing, in part because winds help to disrupt the viscous sublayer at the sea surface, permitting more rapid transport of heat through the surface. In the simplest situation, winds produce a surface stress and a sheared current profile, yielding a classic wall-layer scaling of turbulence and fluxes in the surface layer, similar to the surface layer of the atmosphere. (Theory, supported by experimental observations, predicts a logarithmic velocity profile and constant stress layer in the turbulent layer adjacent to a solid boundary. This is typically found in the atmosphere during neutral stratification and is termed wall-layer scaling.) This simple case, however, seems to be rare. The reason for the difference in behaviors of oceanic and atmospheric surface layers is the difference in the boundaries. The lower boundary of the atmosphere is solid (at least over land, where convection is well-understood), but the ocean's upper boundary is free to support waves, ranging from centimeter-scale capillary waves, through wind waves (10s of meters) to swell (100s of meters). The smaller wind waves lose coherence rapidly, and are therefore governed by local forcing conditions. Swell is considerably more persistent, and may therefore reflect conditions at a location remote in space and time from the observation, e.g., a distant storm.

Breaking Waves

Large scale breaking of waves is evidenced at the surface by whitecapping and surface foam, allowing visual detection from above. This process, which is not at all well understood, disrupts the ocean's cool skin, a fact highlighted by acoustic detection of bubbles injected beneath the sea surface by breaking waves. Small-scale breaking, which has no visible signature (and is even less well understood but is thought to be due to instabilities formed in concert with the superposition of smaller-scale waves) also disrupts the ocean's cool skin. An important challenge for oceanographers is to determine the prevalence of small-scale wave breaking and the statistics of cool skin disruption at the sea surface.

The role of wave breaking in mixing is an issue of great interest at present. Turbulence observations in the surface layer under a variety of conditions have indicated that at times (generally lower winds and simpler wave states) the turbulence dissipation rate (and presumably other turbulence quantities including fluxes) behaves in accordance with simple wall-layer scaling and is in this way similar to the atmospheric surface layer. However, under higher winds, and perhaps more complicated wave states, turbulence dissipation rates greatly exceed those predicted by wall-layer scaling. This condition has been observed to depths of 30 m, well below a significant wave height from the surface, and constituting a significant fraction of the ocean's mixed layer. (The significant wave height is defined as the average height of the highest third of surface displacement maxima. A few meters is generally regarded as a large value.) Evidently, an alternative to wall-layer scaling is needed for these cases. This is a problem of great importance in determining both transfer rates across the air–sea interface to the mixed layer below and the evolution of the mixed layer itself. It is at times when turbulence is most intense that most of the air–sea transfers and most of the mixed layer modification occurs.

Langmuir Circulation

Langmuir circulations are coherent structures within the mixed layer that produce counterrotating vortices with axes aligned parallel to the wind. Their surface signature is familiar as windrows: lines of bubbles and surface debris aligned with the wind that mark the convergence zones between the vortices. These convergence zones are sites of downwind jets in the surface current. They concentrate bubble clouds produced by breaking waves, or bubbles produced by rain, which are then carried downward, enhancing gas-exchange rates with the atmosphere. Acoustical detection of bubbles provides an important method for examining the structure and evolution of Langmuir circulations.

Langmuir circulations appear to be intimately related to the Stokes drift, a small net current parallel to the direction of wave propagation, generated by wave motions. Stokes drift is concentrated at the surface and is thus vertically sheared. Small perturbations in the wind-driven surface current generate vertical vorticity, which is tilted toward the horizontal (downwind) direction by the shear of the Stokes drift. The result of this tilting is a field of counterrotating vortices adjacent to the ocean surface, i.e., Langmuir cells. It is the convergence associated with these vortices that concentrates the wind-driven surface current into jets. Langmuir cells thus grow by a process of positive feedback. Ongoing acceleration of the surface current by the wind, together with convergence of the surface current by the Langmuir cells, provides a continuous source of coherent vertical vorticity (i.e., the jets), which is tilted by the mean shear to reinforce the cells.

Downwelling speeds below the surface convergence have been observed to reach more than $0.2\,\mathrm{m\,s^{-1}}$, comparable to peak downwind horizontal flow speeds. By comparison, the vertical velocity scale associated with convection, $w^* = (J_b^0 D)^{1/3}$ is closer to $0.01\,\mathrm{m\,s^{-1}}$. Upward velocities representing the return flow to the surface appear to be smaller and spread over greater area. Maximum observed velocities are located well below the sea surface but also well above the mixed-layer base. Langmuir circulations are capable of rapidly moving fluid vertically, thereby enhancing and advecting the turbulence necessary to mix the weak near-surface stratification which forms in response to daytime heating. However, this mechanism does not seem to contribute significantly to mixing the base of the deeper mixed layer, which is influenced more by storms and strong cooling events.

In contrast, penetration of the deep mixed layer base during convection (driven by the conversion of potential energy of dense fluid plumes created by surface cooling/evaporation into kinetic energy and turbulence) is believed to be an important means of deepening the mixed layer. So also is inertial shear, as explained next.

Wind-Driven Shear

Wind-driven shear erodes the thermocline at the mixed-layer base. Wind-driven currents often veer with depth due to planetary rotation (cf. the Ekman spiral). Fluctuations in wind speed and direction result in persistent oscillations at near-inertial frequencies. Such oscillations are observed almost everywhere in the upper ocean, and dominate the horizontal velocity component of the internal wave field. Because near-inertial waves dominate the vertical shear, they are believed to be especially important sources of mixing at the base of the mixed layer. In the upper ocean, near-inertial waves are generally assumed to be the result of wind forcing. Rapid diffusion of momentum through the mixed layer tends to concentrate shear at the mixed layer base. This concentration increases the probability of small-scale instabilities. The tendency toward instability is quantified by the Richardson number, $Ri = N^2/S^2$, where $N^2 = -(g/\rho)\cdot d\rho/dz$, represents the stability of the water column, and shear, S, represents an energy source for instability. Small values of Ri ($<1/4$) are associated with Kelvin–Helmholtz instability. Through this instability, the inertial shear is concentrated into discrete vortices (Kelvin–Helmholtz billows) with axes aligned horizontally and perpendicular to the current. Ultimately, the billows overturn and generate small-scale turbulence and mixing. Some of the energy released by the instabilities propagates along the stratified layer as high frequency internal gravity waves. These processes are depicted in **Figure 2**. The mixing of fluid from below the mixed layer by inertial shear contributes to increasing the density of the mixed-layer and to mixed-layer deepening.

Temperature Ramps

Another form of coherent structure in the upper ocean has been observed in both stable and unstable conditions. In the upper few meters temperature ramps, aligned with the wind and marked by horizontal temperature changes of 0.1 K in 0.1 m, indicate the upward transport of cool/warm fluid during stable/unstable conditions. This transport is driven by an instability triggered by the wind and perhaps similar to the Kelvin–Helmholtz instability discussed above. It is not yet clearly understood. Because it brings water of different temperature into close contact with the surface, and also because it causes large lateral gradients, this mechanism appears to be a potentially important factor in near-surface mixing.

Effects of Precipitation

Rainfall on the sea surface can catalyze several important processes that act to both accentuate and reduce upper ocean mixing. Drops falling on the surface disrupt the viscous boundary layer, and may carry air into the water by forming bubbles.

Rain is commonly said to 'knock down the seas.' The evidence for this is the reduction in breaking wave intensity and whitecapping at the sea surface. Smaller waves (<20 cm wavelength) may be damped by subsurface turbulence as heavy rainfall acts to transport momentum vertically, causing drag on the waves. The reduced roughness of the small-scale waves reduces the probability of the waves exciting flow separation on the crests of the long waves, and hence reduces the tendency of the long waves to break.

While storm winds generate intense turbulence near the surface, associated rainfall can confine this turbulence to the upper few meters, effectively insulating the water below from surface forcing. This is due to the low density of fresh rainwater relative to the saltier ocean water. Turbulence must work against gravity to mix the surface water downward, and turbulent mixing is therefore suppressed. So long as vertical mixing is inhibited, fluid heated during the day will be trapped near the sea surface. Preexisting turbulence below the surface will continue to mix fluid in the absence of direct surface forcing, until it decays due to viscous dissipation plus mixing, typically over the time scale of a buoyancy period, N^{-1}.

Deposition of pools of fresh water on the sea surface, such as occurs during small-scale squalls, raises some interesting prospects for both lateral spreading and vertical mixing of the fresh water. In the warm pool area of the western equatorial Pacific, intense squalls are common. Fresh light puddles at the surface cause the surface density field to be heterogeneous. Release of the density gradient may then occur as an internal bore forming on the surface density anomaly, causing a lateral spreading of the fresh puddle. Highly resolved horizontal profiles of temperature, salinity and density reveal sharp frontal interfaces, the features of which depend on the direction of the winds relative to the buoyancy-driven current. These are portrayed in **Figure 3**. When the wind opposes the buoyancy current, the density anomaly at the surface is reduced, possibly as a result of vertical mixing in the manner suggested in **Figure 3(B)**. This mechanism results in a rapid vertical redistribution of fresh water from the surface pool and a brake on the propagation of the buoyancy front. Similarly, an opposing ambient current results in shear at the base of the fresh layer, which may lead to instability and consequent mixing. The nature of these features has yet to be clearly established, as has the net effect on upper ocean mixing.

Ice on the Upper Ocean

At high latitudes, the presence of an ice layer (up to a few meters thick) partially insulates the ocean against surface forcing. This attenuates the effects of wind forcing on the upper ocean except at the lowest frequencies. The absence of surface waves prevents turbulence due to wave breaking and Langmuir circulation. However, a turbulence source is

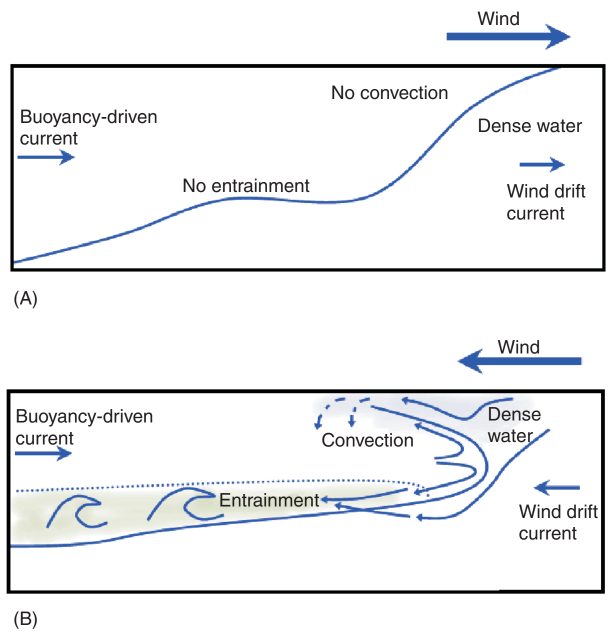

Figure 3 Two ways in which the frontal interface of a fresh surface pool may interact with ambient winds and currents. (A) The case in which the buoyancy-driven current, wind and ambient current are all in the same direction. In this case, the buoyancy-driven current spreads and thins unabated. In (B), the buoyancy-driven current is opposed by wind and ambient current. In this case, the frontal interface of the buoyancy-driven current may plunge below the ambient dense water, so that convection near the surface intensifies mixing at the frontal interface. Simultaneously, shear-forced mixing at the base of the fresh puddle may increase entrainment of dense water from below.

provided by the various topographic features found on the underside of the ice layer. These range in size from millimeter-scale dendritic structures to 10 m keels, and can generate significant mixing near the surface when the wind moves the ice relative to the water below or currents flow beneath the ice.

Latent heat transfer associated with melting and freezing exerts a strong effect on the thermal structure of the upper ocean. Strong convection can occur under ice-free regions, in which the water surface is fully exposed to cooling and evaporative salinity increase. Such regions include leads (formed by diverging ice flow) and polynyas (where wind or currents remove ice as rapidly as it freezes). Convection can also be caused by the rejection of salt by newly formed ice, leaving dense, salty water near the surface.

Parameterizations of Upper Ocean Mixing

Large-scale ocean and climate models are incapable of explicitly resolving the complex physics of the upper ocean, and will remain so for the foreseeable future. Since upper ocean processes are crucial in determining atmosphere–ocean fluxes, methods for their representation in large-scale models, i.e., parameterizations, are needed. The development of upper-ocean mixing parameterizations has drawn on extensive experience in the more general problem of turbulence modeling. Some parameterizations emphasize generality, working from first principles as much as possible, whereas others sacrifice generality to focus on properties specific to the upper ocean. An assumption common to all parameterizations presently in use is that the upper ocean is horizontally homogeneous, i.e., the goal is to represent vertical fluxes in terms of vertical variations in ocean structure, leaving horizontal fluxes to be handled by other methods. Such parameterizations are referred to as 'one-dimensional' or 'column' models.

Column modeling methods may be classified as local or nonlocal. In a local method, turbulent fluxes at a given depth are represented as functions of water column properties at that depth. For example, entrainment at the mixed-layer base may be determined solely by the local shear and stratification. Nonlocal methods allow fluxes to be influenced directly by remote events. For example, during nighttime convection, entrainment at the mixed-layer base may be influenced directly by changes in the surface cooling rate. In this case, the fact that large convection rolls cannot be represented explicitly in a column model necessitates the nonlocal approach. Nonlocal methods include 'slab' models, in which currents and water properties do not vary at all across the depth of the mixed layer. Local representations may often be derived systematically from the equations of motion, whereas nonlocal methods tend to be *ad hoc* expressions of empirical knowledge. The most successful models combine local and nonlocal approaches.

Many processes are now reasonably well represented in upper ocean models. For example, entrainment via shear instability is parameterized using the local gradient Richardson number and/or a nonlocal (bulk) Richardson number pertaining to the whole mixed layer. Other modeling issues are subjects of intensive research. Nonlocal representations of heat fluxes have resulted in improved handling of nighttime convection, but the corresponding momentum fluxes have not yet been represented. Perhaps the most important problem at present is the representation of surface wave effects. Local methods are able to describe the transmission of turbulent kinetic energy generated at the surface into the ocean interior. However, the dependence of that energy flux on surface forcing is complex and remains poorly understood. Current research into the physics of wave breaking, Langmuir circulation, wave-precipitation interactions, and other surface wave phenomena will lead to improved understanding, and ultimately to useful parameterizations.

See also

Breaking Waves and Near-surface Turbulence. Bubbles. Deep Convection. Heat and Momentum Fluxes at the Sea Surface. Internal Tides. Langmuir Circulation and Instability. Penetrating Shortwave Radiation. Surface, Gravity and Capillary Waves. Three-dimensional (3D) Turbulence. Under-ice Boundary Layer. Upper Ocean Vertical Structure. Whitecaps and Foam. Wave Generation by Wind.

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UPPER OCEAN RESPONSES TO STRONG FORCING EVENTS

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Introduction

Strong atmospheric forcing events occur over the oceans during both summer and winter months. Cyclonically (anticyclonically) rotating wind fields around a surface low-pressure region in the Northern (Southern) Hemisphere excite an energetic ocean response. For example, winter storms and extratropical cyclones originating in the Gulf of Alaska (north-east Pacific Ocean) or over the Gulf Stream (western Atlantic Ocean) significantly impact the north-west and north-east coasts of the United States, respectively. Similarly, tropical storm formation and their subsequent development into tropical cyclones (called hurricanes in the Eastern Pacific and Atlantic Ocean basins, and typhoons in the Western Pacific Ocean) cause an energetic ocean response in these basins. In this context, the ocean's response is pronounced as manifested in the upper ocean cooling patterns that are modulated by the three-dimensional current structure excited by storms.

Observationally, studies over the past decade have demonstrated the importance of the current structure on the oceanic mixed layer (OML) thermal response. Thermal structure changes using temperature profiles and remotely sensed data acquired during hurricane conditions have been well documented. However, the oceanic current response to the surface winds has focused primarily on analytical and numerical solutions with simplifying assumptions. A key component of the current response is associated with the divergence and convergence of the OML currents that induce upwelling and downwelling regimes, respectively, in the thermocline. The wind-forced currents rotate anticyclonically with time with a period of oscillation close to the local inertial period. In addition, this wind-forced, near-inertial current vector rotates anticyclonically with depth and creates significant shear across the OML base that causes vertical mixing and upper ocean cooling. Although the current response has been thought to be confined to the upper part of the water column, fortuitous encounters of hurricanes with spatially limited current meter moorings indicate that the response extends through the thermocline to as deep as 1000 m. These data have provided a new view of the oceanic current response to storms, and have challenged theories and models concerning strongly forced conditions.

Accordingly, the ocean's current response encompasses both the directly forced or near-field region, and the evolving three-dimensional wake or far-field regime. In the near-field region, the cyclonically rotating wind field of a tropical cyclone forces the OML currents of about 1–$2\,\mathrm{m\,s^{-1}}$ to diverge from the storm track starting within one-quarter of an inertial wavelength (Λ) behind the eye, defined as the product of the storm translation speed U_h and the local inertial period (IP). This OML current divergence and net Ekman transport away from the storm track cause the upwelling of cooler water that decreases the OML depth (**Figure 1**). Over the next half of a near-inertial cycle, OML currents converge toward the storm track, causing an increase in the OML depth as warmer water is downwelled into the thermocline. This alternating cycle of upwelling and downwelling of the isotherms occurs over