measurements from the NOAA WP-3D in August 1999 indicates a striking contrast between the WCR and Loop Current and the Gulf common water (**Figure 12**). In this deep warm reservoir, the isotherm depth of the 26°C water is 120–130 m where heat content is a maximum of 130 kJ cm^{-2}. This is a factor of 8–10 times larger than is necessary to maintain a tropical cyclone. Notice that the WCR north west of the Loop Current also contains the same oceanic characteristics. By contrast, the Gulf common water (lower left) has isotherm depths of 40 m with heat content of less than half that of the Loop Current and WCR system. This implies that the warm frontal boundary currents and rings will influence the air–sea fluxes feeding the storm that may cause a storm to intensify if atmospheric conditions are favorable. In addition to using airborne measurements, this area of research also utilizes satellite-based radar altimeters in assessing and monitoring upper ocean heat content. From a practical standpoint, accurate monitoring of these processes and the ensuing air–sea fluxes are crucial to improve forecasts of storm intensity within 36 h of landfall. Thus, this new avenue of research has applications to the operational community in providing forecasts for landfalling storms where warm subtropical water is located close to the coast.

See also

Expendable Sensors. Heat and Momentum Fluxes at the Sea Surface. Ocean Circulation. Wind Driven Circulation.

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UPPER OCEAN TIME AND SPACE VARIABILITY

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Introduction

The upper ocean is the region of the ocean in direct contact with the atmosphere. Air–sea fluxes of momentum, heat, and fresh water are the primary external forces acting upon the upper ocean (*see* **Heat and Momentum Fluxes at the Sea Surface; Evaporation and Humidity; and Wind and Buoyancy-forced Upper Ocean**). These fluxes impose the temporal and spatial scales of the overlying atmosphere. The internal dynamics of the ocean cause variability at scales distinct from the forcing. This combination of forcing and dynamics creates the tapestry of oceanic phenomena at timescales ranging from minutes to decades and length scales from centimeters to thousands of kilometers.

This article is concerned primarily with the physical processes causing time and space variability in the upper ocean. The physical balances to be considered are the conservation of mass, heat, salt, and momentum. Thus, physical phenomena are discussed with special reference to their effects on the temporal and spatial variability of temperature, salinity, density, and velocity. While many other biological, chemical, and optical properties of the ocean are affected by the phenomena outlined below, their discussion is covered by other articles in this volume.

The most striking feature often seen in vertical profiles of the upper ocean is the surface mixed layer, a layer that is vertically uniform in temperature, salinity, and horizontal velocity (*see* **Upper Ocean Vertical Structure** and **Upper Ocean Mean Horizontal Structure**). The turbulence that mixes this layer derives its energy from wind and surface cooling. The region immediately below the mixed layer tends to be stratified, and is often called the seasonal thermocline because its stratification varies with the seasons. The seasonal thermocline extends down a few hundred meters to roughly 1000 m. Beneath the seasonal thermocline is the permanent thermocline whose stratification is constant on timescales of at least decades. Here the discussion is concerned with variability of the mixed layer and seasonal thermocline.

The processes discussed below are ordered roughly by increasing time and space scales (**Figure 1**). Most of the processes are covered in greater detail elsewhere in this volume. It is hoped that this section will provide a convenient introduction to the variability of the upper ocean, and that the reader can proceed to the more in-depth articles as needed.

Turbulence and Mixing

The upper ocean is distinguished from the interior of the ocean partly because of the very high levels of turbulence present (*see* **Breaking Waves and**

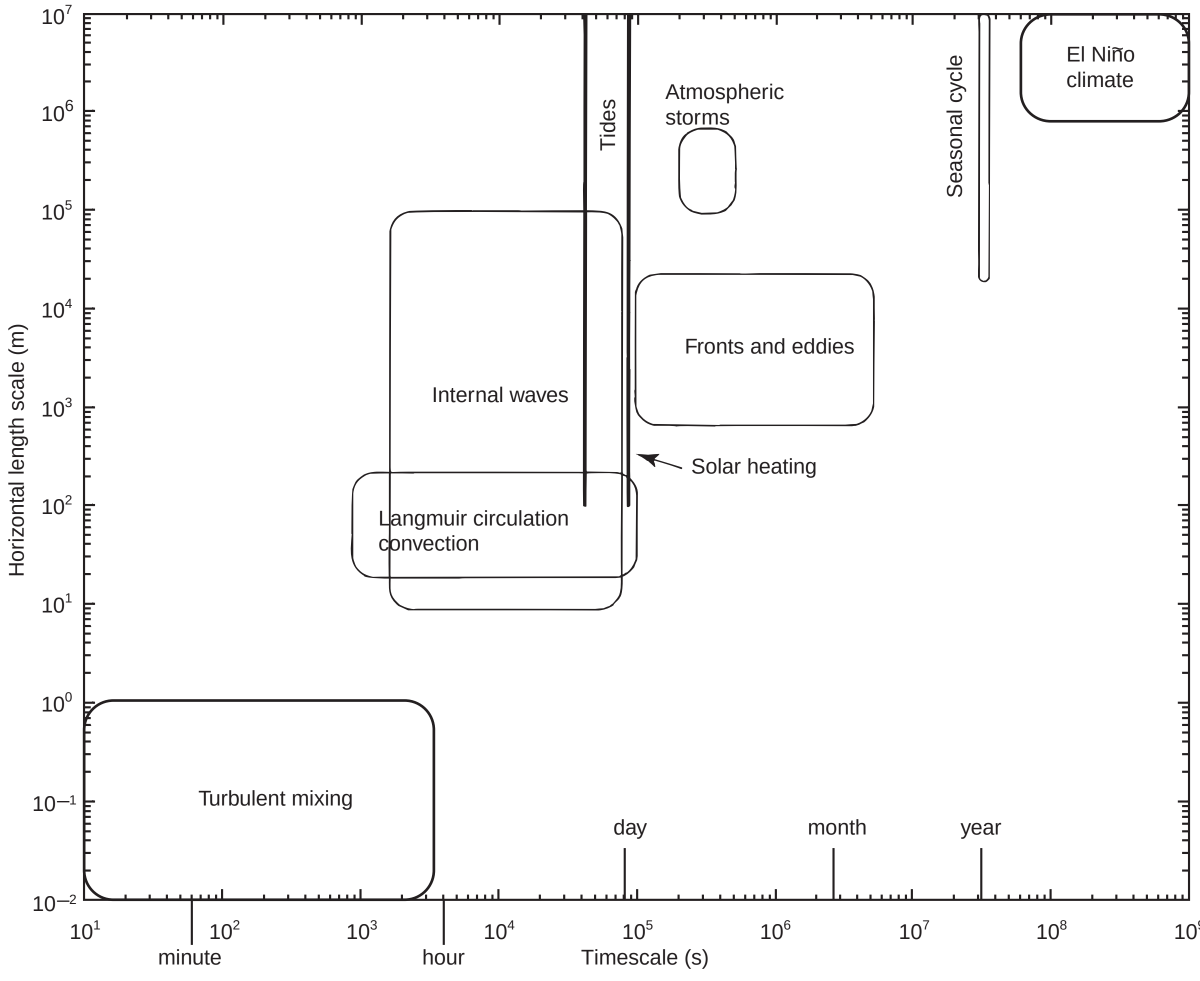

Figure 1 A schematic diagram of the distribution in time and space of upper ocean variability. The temporal and spatial limits of the phenomena should be considered approximate.

Near-surface Turbulence and **Upper Ocean Mixing Processes**). The smallest scale of motion worthy of note in the ocean is the Kolmogoroff scale, on the order of 1 cm, where energy is dissipated by molecular viscosity. At this scale, the ocean can be considered isotropic; that is, properties vary in the same way regardless of the direction in which they are measured. At much larger scales than the Kolmogoroff scale, the vertical stratification of the ocean becomes important.

In the seasonal thermocline, a dominant mechanism for mixing is the Kelvin-Helmholtz instability, in which a vertical shear of horizontal velocity causes the overturn of stratified water (*see* **Internal Waves**). The resulting 'billows' are observed to be on the order of 1 m thick and to decay on the order of an hour. A great deal of observational and theoretical work in the last 20 years has been devoted to relating the strength of this mixing to larger (in the order of 10 m) and more easily measurable quantities such as shear and stratification. The resulting Henyey-Gregg parameterization is one of the most fundamentally important achievements of modern oceanography.

Langmuir Circulation and Convection

Turbulence in the mixed layer is fundamentally different from that in the seasonal thermocline. Because the mixed layer is nearly unstratified, the largest eddies can be as large as the layer is thick, often about 100 m. These large eddies have come to be called Langmuir cells in honor of Irving Langmuir, the Nobel laureate in chemistry who first described them. Langmuir cells are elongated vortices whose axes are horizontal and oriented nearly parallel to the wind. The cells have radii comparable in size to the mixed layer depth, and can be as long as 1–2 km. Langmuir cells often appear in pairs with opposite senses of rotation. The cells thus create alternating regions of surface convergence and divergence. The regions of convergence collect material floating on the surface such as oil and seaweed. Langmuir first became aware of these cells after noticing lines of floating seaweed during a crossing of the Atlantic. Langmuir cells are forced by a combination of wind and surface waves, and are established typically within an hour after the wind starts blowing. Langmuir cells disappear quickly after the wind stops. Recent research indicates that Langmuir cells often vacillate in strength on the timescale of roughly 15 minutes.

Convection cells forced by surface cooling also cause the mixed layer to be homogenized and to deepen (*see* **Open Ocean Convection**). A typical feature in the mixed layer is the daily cycle of stratification, with daytime heating causing near-surface stratification and nighttime cooling causing convection that destroys this stratification and deepens the mixed layer. The vertical extent of convection cells corresponds to the depth of the mixed layer (of order 100 m); the cells have an aspect ratio of one so their horizontal and vertical scales are equal. Because solar heating has a large, essentially global, scale the daily heating and cooling of the upper ocean is coherent and predictable over large scales. Horizontal velocity in the mixed layer also varies strongly at a 24 h period, as the daily cycle of stratification affects the depth to which the wind forces currents. The deepest mixed layers in the oceans, at high latitudes, are convectively mixed. Convection cells are thus more effective at deepening the mixed layer than are Langmuir cells.

Internal Waves

Just as there are gravity waves on the surface of the ocean, there are gravity waves in the thermocline. These thermocline gravity waves, modified by the Earth's rotation, are known as internal waves (*see* **Internal Waves**). They exist in a range of frequencies bounded at the lower end by the inertial frequency f and at the upper end by the buoyancy frequency N. A parcel of water given an initial velocity will travel in a circle under the influence of the Coriolis force. The inertial frequency f, twice the local vertical component of the Earth's rotation vector, is the frequency of rotation around such a circle. The resulting horizontal current is known as an inertial oscillation. The inertial period is 12 h at the poles, 24 h at 30° latitude, and infinite at the equator because local vertical is normal to the Earth's axis of rotation. The buoyancy frequency N, proportional to the square root of the vertical density gradient, is the frequency of oscillation of a water parcel given a displacement in the vertical. The resulting vertical motion has a frequency of less than one to several cycles per hour in typical ocean stratification. Internal waves oscillate in planes tilted from the horizontal as a function of the frequency between f and N. Internal waves have amplitudes on the order of tens of meters. They may be coherent over vertical scales that approach the depth of the ocean, particularly at high frequencies near N. Lower frequency internal waves, approaching f, have shorter vertical wavelengths often of order 100 m or less. The horizontal wavelength of an internal wave is related to its frequency and vertical wavelength through the internal wave dispersion relation. For a given vertical wavelength, a high frequency internal wave will have shorter horizontal wavelength than a low frequency wave.

At the low frequency end of the internal wave spectrum, the near-inertial waves are especially important in the upper ocean. Near-inertial waves are quite ubiquitous because they are so readily excited by wind forcing on the ocean's surface. In measurements of horizontal current, inertial oscillations are often the most obvious variability because horizontal currents 'ring' at the resonant inertial frequency. Just as a bell has a distinctive tone when struck, the ocean has inertial currents when hit, for example, by a storm. Strong inertial currents are one of the indications in the ocean of the recent passage of a hurricane. The radius of an inertial current circle is its speed divided by its rotation rate, U/f. If the current speed is $0.1\,\mathrm{m\,s^{-1}}$, then for a midlatitude inertial frequency of $10^{-4}\,\mathrm{s^{-1}}$, the radius is 1 km. In the aftermath of a storm, the inertial currents and radii may be nearly an order of magnitude larger. Near-inertial waves are a dominant mechanism for transporting wind-driven momentum downward from the mixed layer to the seasonal thermocline and into the interior. Because near-inertial motions have short vertical scales, they dominate the shear spectrum in the ocean. This shear eventually leads to enhanced turbulence and mixing the penetration of inertial shear into the ocean and the geography of shear and mixing are active topics of research.

Tides are well known to anyone who has spent at least a day at the beach. The dominant tidal periods are near one day and one-half day. Tides are most obvious to the casual observer of the sea surface, and they are easily seen in records of horizontal current in the open ocean. Internal tides exist as well, for example forced by tidal flow over bumps on the ocean bottom (*see* **Internal Waves**). These internal tides, seen as variability in density and velocity at a location, are a form of internal wave and are governed by the same dynamics. Isolated pulses of tidal internal waves, known as 'solitons', are prevalent in certain regions of rough bottom topography, and are a field of current research.

Fronts and Eddies

While vertically uniform, the mixed layer can vary in the horizontal on a wide range of scales. We have already discussed Langmuir circulation and convection cells on scales of order 100 m, but there may be horizontal variability on longer scales. Just as there are fronts in the atmosphere, visible for example in the satellite pictures of clouds shown on the evening television news, there are fronts in the ocean. Fronts in the ocean separate regions of warm and cool water, or fresh and salty water. The most obvious fronts in the mixed layer have widths on the order of 10–100 km, and typically persist for weeks. Fronts of this size have currents directed along the front as a result of the geostrophic momentum balance. That is, the Coriolis force balances the pressure gradient due to having water of varying density across the front. The less dense (usually warmer) water is on the right side of the current in the Northern Hemisphere (the sense of the current is the opposite in the Southern Hemisphere). Fronts in the mixed layer are sites of enhanced vertical circulation on the order of tens of meters per day. Strong biological productivity at fronts is attributed to this vertical circulation which brings deeper water rich in nutrients to the surface.

Fronts at scales shorter than 10 km also exist in the mixed layer. At these shorter scales, the geostrophic balance may not be expected to hold. Typical fronts at these scales are observed to be warm and salty on one side and cold and fresh on the other such that the density contrast across the front vanishes. Such a front is often said to be compensated, since temperature and salinity gradients compensate in their effect on density. The presence of compensated fronts in the mixed layer is consistent with a horizontal mixing that is an increasing function of the horizontal density gradient. That is, small-scale horizontal density fronts do not persist as long as compensated fronts. Because of their small scale, fronts of order 1 km are poorly observed in the ocean, and are a topic of current research.

Observed fronts are usually not observed to be perfectly straight, rather they wiggle. The wiggles, or perturbations, often grow to be large in comparison with the width of the front. When the perturbations grow large enough, the front may turn back on itself and a detached eddy is formed. The eddies often have sizes on the order of 10 km, when they are confined in depth to the mixed layer. This length scale is related to the Rossby radius of deformation; at scales larger than the Rossby radius flows tend to be geostrophic. The Rossby radius for the mixed layer is given by:

$$\frac{\sqrt{gH\Delta\rho/\rho}}{f}$$

where g is acceleration due to gravity, H is the depth of the mixed layer, ρ is the density of the water, and $\Delta\rho$ is the change in density across the mixed layer base. For a typical mixed layer, H is

100 m and $\Delta\rho$ is $0.2\,\mathrm{kg\,m^{-3}}$, g is $9.8\,\mathrm{m\,s^{-2}}$, and ρ is $1025\,\mathrm{kg\,m^{-3}}$, so the Rossby radius is about 6 km. Eddies that extend deeper have larger radii, as can be inferred from the formula for the Rossby radius. Large eddies can persist for as long as several months, while smaller eddies are shorter lived. The small-scale mixed layer eddies, a prominent feature in satellite photos of the sea surface, are typically observed to rotate in the counterclockwise direction in the Northern Hemisphere, and clockwise south of the equator. Again, because of their small size, they have been inadequately observed and are a topic of current research.

Wind-Forced Currents (*see* Wind Driven Circulation)

One of the oldest theories of ocean circulation is due to V.W. Ekman, who in 1905 suggested a balance between the Coriolis force and the stress due to wind blowing over the ocean surface. The prediction of this theory for a steady wind is a current that spirals to the right (in the Northern Hemisphere) and decays with depth. This spiral structure was not clearly observed in the ocean until the 1980s with the advent of moorings with modern current meters. Although the details of the stress parameterization used by Ekman were found to be inadequate to describe observations, the general picture of a spiral remains valid to this day.

An alternative theoretical construct to explain upper ocean structure is the bulk mixed layer model. Oceanic properties, such as temperature, salinity, and velocity, are assumed to be vertically uniform in the mixed layer, with a region of very strong vertical gradients at the mixed layer base. The mixed layer is then forced by air–sea fluxes of heat, fresh water, and momentum at the surface, and by turbulent fluxes at the base. The bulk mixed layer model has proven remarkably successful at predicting some basic features of the upper ocean, particularly the vertical temperature structure.

Interestingly, the disparate conceptual models of the Ekman spiral and the bulk mixed layer can be rationalized. The upper ocean velocity structure is often, but certainly not always, observed to be vertically uniform near the surface with a region of high shear beneath, in accordance with the bulk mixed layer model. On the other hand, long time averages of ocean current tend to have a spiral structure, in qualitative agreement with the Ekman spiral. This is so if the averages are long enough to span many cycles of mixed layer shoaling and deepening, as due to the daily cycle of surface heating. Thus the time-average current spiral may be very different from a typical snapshot of a nearly vertically uniform current.

The averaged wind-driven spiral extends downward to a depth comparable to, but slightly deeper than, the mixed layer. The shape of the spiral is strongly influenced by higher frequency variability in the stratification, such as the daily cycle in mixed layer depth discussed above. A spiral is observed in response to temporally variable winds, as well as to steady winds. The temporally variable spiral may have a different vertical structure to the steady spiral. In particular, the current spirals to the left with depth in response to a wind that rotates more rapidly than f in a clockwise direction, in contrast to the steady spiral to the right.

Regardless of the detailed velocity structure in the upper ocean, the net transport caused by a steady wind is 90° to the right of the wind in the Northern Hemisphere (and to the left in the Southern Hemisphere). This transport (the vertical integral of velocity) is called the Ekman transport. The Ekman transport is proportional to the wind stress and inversely proportional to the inertial frequency. Thus wind of a given strength will cause more transport near the equator than it would closer to the poles.

The spectrum of wind over the midlatitude ocean peaks at periods of a few to several days. These periods correspond to the time required for a typical storm to pass. The wind-driven current and transport is thus prominent at these periods. Atmospheric storms have typical horizontal sizes of a few to several hundred kilometers, and the direct oceanic response to these storms has similar horizontal scales. The prominent large-scale features of the wind field such as the westerlies in midlatitudes and the trade winds in the tropics directly force currents in the upper ocean. These currents have large horizontal length scales that reflect the winds.

Seasonal Cycles

Just as the seasons cause well-known changes in weather, the annual cycle is one of the most robust signals in the ocean. Summer brings greater heat flux from the atmosphere to the ocean, and warmer ocean temperatures. As the ocean warms up at the surface, stratification increases and the mixed layer becomes shallower. The heat flux reverses in many locations during the winter and the ocean cools at the surface. The resulting convection causes the mixed layer to deepen; at some high latitude locations the mixed layer can deepen to several hundred meters in the winter. Winter conditions in high and midlatitude mixed layers are very important to the general circulation of the oceans, as it is these waters that penetrate into the thermocline and set properties that persist for decades. Along with cooler temperatures, winter brings typically stormier weather and more wind and precipitation. Wind-driven currents often peak during the winter in mid-latitudes, at the same time that salinity decreases in response to the increased precipitation.

Seasonal cycles occur over the whole globe in an extremely coherent fashion, because they are driven primarily by the solar heat flux. However, the seasonal cycle can vary at different oceanic locations. For example, the seasonal cycle at the equator is smaller than that at midlatitudes because solar heat flux varies less over the year. The Arabian Sea has a pronounced semi-annual cycle. Cold northerly winds in winter cool the ocean and deepen the mixed layer as typical for midlatitudes. More unusual is a second period of relatively low ocean temperatures and deep mixed layers during the summer south-west monsoon. Wind-driven mixing causes the cooling during the south-west monsoon as cool water is mixed up to the surface. The Arabian Sea monsoon is the classic example of a seasonal wind driven by land–sea temperature differences. Monsoons also exist over the south-west USA and south-east Asia, among others. Additional local seasonal effects may be caused by river outflows and weather patterns influenced by orography.

Climatic Signals

The ocean has significant variability at periods longer than 1 year. The most well known recurrent interannual climatic phenomenon is El Niño (*see* **El Niño Southern Oscillation (ENSO)**). An El Niño occurs when trade winds reverse at the equator causing upwelling to cease off the coast of South America. The most obvious consequence of an El Niño is dramatically elevated ocean temperatures at the equator. These high temperatures progress poleward from the equator along the coast of the Americas, affecting water properties in large regions of the Pacific. El Niño has been hypothesized to start with anomalous winds in the western equatorial Pacific, eventually having an effect on the global ocean and atmosphere. El Niños occur sporadically every roughly 3–7 years, and are becoming more predictable as observations and models of the phenomenon improve. The reverse phase of El Niño, the so-called La Nina, is remarkable for exceptionally low equatorial temperatures and strong trade winds.

Oscillations with periods of a decade and longer also exist in the ocean and atmosphere. Such oscillations are apparent in the ocean as basin-scale variations in sea surface temperature, for example. Salinity and velocity are also likely variable on decadal timescales, although the observational database for these is sparse in comparison with that for temperature. Atmospheric decadal oscillations in temperature and precipitation are well established. Scientists are actively researching whether and how the ocean and atmosphere are coupled on decadal timescales. The basic idea is that the ocean absorbs heat from the atmosphere and stores it for many years because of the ocean's relatively high heat capacity. This heat may penetrate into the ocean interior and be redistributed by advective processes. The heat may resurface a decade or more later to affect the atmosphere through anomalous heat flux. The coupled ocean–atmosphere process just described is controversial, and the observations to support its existence are inadequate. A major challenge for the immediate future is to obtain the measurements needed to resolve such processes of significance to climate.

Conclusion

The upper ocean varies on a wide range of temporal and spatial scales. Processes range from mixing occurring on scales of centimeters and minutes to decadal climatic oscillations of entire ocean basins. Fundamental to the ocean is the fact that these processes can rarely be studied in isolation. That is, processes occurring on one scale affect processes on other scales. For example, decadal changes in ocean stratification are strongly affected by turbulent mixing at the smallest scales. Turbulent mixing is modulated by the internal wave field, and internal waves are focused and steered by geostrophic fronts and eddies. The interaction among processes of different scales is likely to receive increasing attention from ocean scientists in the coming years.

See also

Breaking Waves and Near-surface Turbulence. Double-diffusive Convection. El Niño Southern Oscillation (ENSO). Evaporation and Humidity. Heat and Momentum Fluxes at the Sea Surface. Internal Waves. Open Ocean Convection. Upper Ocean Mean Horizontal Structure. Upper Ocean Mixing Processes. Upper Ocean Vertical Structure. Wind and Buoyancy-forced Upper Ocean. Wind Driven Circulation.

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UPPER OCEAN VERTICAL STRUCTURE

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Introduction

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

Major Features of the Upper Ocean Vertical Structure

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

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