h^{-1}), only a small fraction of the sea surface is covered by stage B whitecaps (0.04 or 4%), and an even smaller fraction of that surface is covered by stage A whitecaps (0.002 or 0.2%). Yet the total area of all the world's oceans is very great ($3.61 \times 10^{14}\,m^2$), and as a consequence the total area of the global ocean covered by whitecaps at any instant is considerable. If a wind speed of $7\,m\,s^{-1}$ is taken as a representative value, then at any instant some $7.0 \times 10^{10}\,m^2$, i.e. some $70\,000\,km^2$, of stage A whitecap area is present on the surface of the global ocean. Following from this, and including such additional information as the terminal rise velocity of bubbles, it can be deduced that some $7.2 \times 10^{11}\,m^2$, i.e. some $720\,000\,km^2$ of individual bubble surface area are destroyed each second in all the stage A whitecaps present on the surface of all the oceans, and an equal area of bubble surface is being generated in the same interval. The vast amount of bubble surface area destroyed each second on the surface of all the world's oceans, and the great volume of water (some $2.5 \times 10^{11}\,m^3$) swept by all the bubbles that burst on the sea surface each second, have profound implications for the global rate of air–sea exchange of moisture, heat and gases. An additional preliminary calculation following along these lines, suggests that all the bubbles breaking on the sea surface each year collect some 2 Gt of carbon during their rise to the ocean surface.

See also

Heat and Momentum Fluxes at the Sea Surface. Wave Generation by Wind.

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WIND AND BUOYANCY-FORCED UPPER OCEAN

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Introduction

0001   Forcing from winds, heating and cooling, and rainfall and evaporation, have a profound influence on the distribution of mass and momentum in the ocean. Although the effects from this wind and buoyancy forcing are ultimately felt throughout the entire ocean, the most immediate impact is in the surface mixed layer, the site of the active air–sea exchanges. The mixed layer is warmed by sunshine and cooled by radiation emitted from the surface and by latent heat loss due to evaporation (**Figure 1**). The mixed layer also tends to be cooled by sensible heat loss since the surface air temperature is generally cooler than the ocean surface. Evaporation and precipitation change the mixed layer salinity. These salinity and temperature changes define the ocean's surface buoyancy. As the surface loses buoyancy, the surface can become denser than the subsurface waters, causing convective overturning and

mixing to occur. Wind forcing can also cause surface overturning and mixing, as well as localized overturning at the base of the mixed layer through sheared-flow instability. This wind- and buoyancy-generated turbulence causes the surface water to be well mixed and vertically uniform in temperature, salinity, and density. Furthermore, the turbulence can entrain deeper water into the surface mixed layer, causing the surface temperature and salinity to change and the layer of well-mixed, vertically uniform water to thicken. Wind forcing can also set up oceanic currents and cause changes in the mixed layer temperature and salinity through horizontal and vertical advection.

0002   Although the ocean is forced by the atmosphere, the atmosphere can also respond to ocean surface conditions, particularly sea surface temperature (SST). Direct thermal circulation, in which moist air rises over warm SSTs and descends over cool SSTs, is most prevalent in the tropics. The resulting atmospheric circulation cells influence the patterns of cloud, rain, and winds that combine to form the wind and buoyancy forcing for the ocean. Thus, the oceans and atmosphere form a coupled system,

where it is sometimes difficult to distinguish forcing from response. Because water has a heat capacity and density nearly three orders of magnitude larger than air, the ocean has thermal and mechanical inertia relative to the atmosphere. The ocean thus acts as a memory for the coupled ocean–atmosphere system.

We begin with a discussion of air–sea interaction   0003 through surface heat fluxes, moisture fluxes, and wind forcing. The primary external force driving the ocean–atmosphere system is radiative warming from the sun. Because of the fundamental importance of solar radiation, the surface wind and buoyancy forcing is illustrated here with two examples of the seasonal cycle. The first case describes the seasonal cycle in the north Pacific, and can be considered a classic example of a one-dimensional (involving only vertical processes) ocean response to wind and buoyancy forcing. In the second example, the seasonal cycle of the eastern tropical Pacific, the atmosphere and ocean are coupled, so that wind and buoyancy forcing lead to a sequence of events that make cause and effect difficult to determine. The impact of wind and buoyancy forcing on the surface
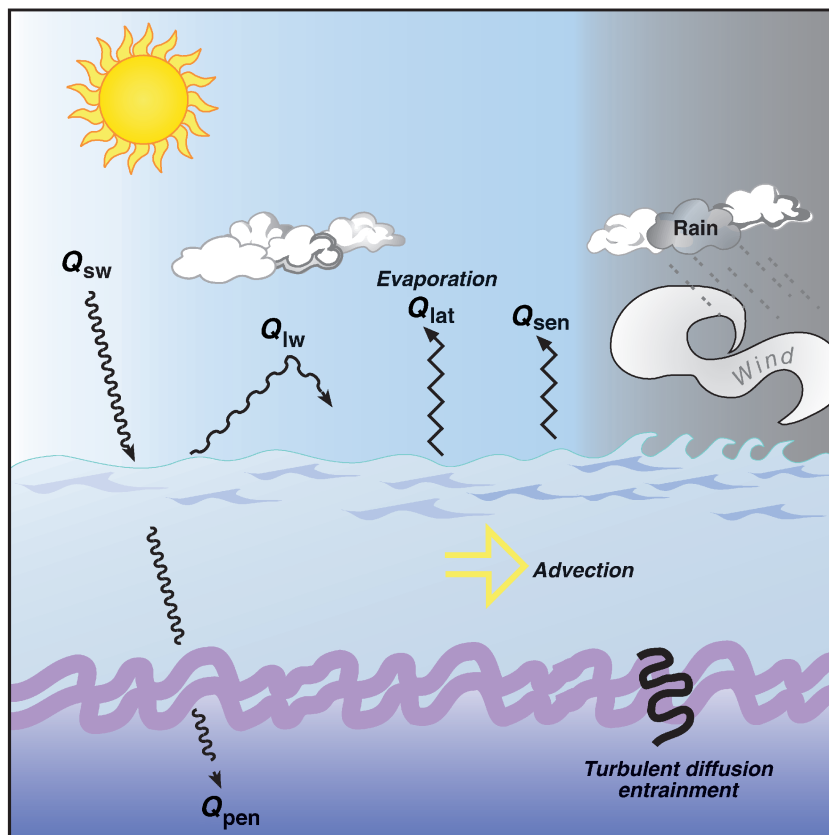
a0157f0001   **Figure 1**   Wind and buoyancy forces acting on the upper ocean mixed layer. Solar radiation (Q_{sw}), net longwave radiation (Q_{lw}), latent heat flux (Q_{lat}), and sensible heat flux (Q_{sen}) combine to form the net surface heat flux (Q_0). Q_{pen} is the solar radiation penetrating the base of the mixed layer.

mixed layer and the deeper ocean is summarized in the conclusion.

Air–Sea Interaction

Surface Heat Flux

0004 As shown in **Figure 1**, the net surface heat flux entering the ocean (Q_0) includes solar radiation (Q_{sw}), net long-wave radiation (Q_{lw}), latent heat flux due to evaporation (Q_{lat}), and sensible heat flux due to air and water having different surface temperatures (Q_{sen}):

$$Q_0 = Q_{sw} + Q_{lw} + Q_{lat} + Q_{sen} \qquad [1]$$

The Earth's seasons are largely defined by the annual cycle in the net surface heat flux associated with the astronomical orientation of the Earth relative to the Sun. The Earth's tilt causes solar radiation to strike the winter hemisphere more obliquely than the summer hemisphere. As the Earth orbits the sun, winter shifts to summer and summer shifts to winter, with the sun directly overhead at the equator twice per year, in March and again in September. Thus, one might expect the seasonal cycle in the tropics to be semiannual, rather than annual. However, as discussed later, in some parts of the equatorial oceans, the annual cycle dominates due to coupled ocean–atmosphere–land interactions.

0005 Solar radiation entering the Earth's atmosphere is absorbed, scattered, and reflected by water in both its liquid and vapor forms. Consequently, the amount of solar radiation which crosses the ocean surface, Q_{sw}, also depends on the cloud structures. The amount of solar radiation absorbed by the ocean mixed layer depends on the transmission properties of the light in the water and can be estimated as the difference between the solar radiation entering the surface and the solar radiation penetrating through the base of the mixed layer (i.e., $Q_{sw} - Q_{pen}$ in **Figure 1**).

0006 The Earth's surface also radiates energy at longer wavelengths similar to a black-body (i.e., proportional to the fourth power of the surface temperature in units kelvin). Long-wave radiation emitted from the ocean surface can reflect against clouds and become downwelling long-wave radiation to be reabsorbed by the ocean. Thus net long-wave radiation, Q_{lw}, is the combination of the outgoing and incoming long-wave radiation and tends to cool the ocean.

0007 The ocean and atmosphere also exchange heat via conduction ('sensible' heat flux). When the ocean and atmosphere have different surface temperatures, sensible heat flux will act to reduce the temperature gradient. Thus when the ocean is warmer than the air (which is nearly always the case), sensible heat flux will tend to cool the ocean and warm the atmosphere. Likewise, the vapor pressure at the air–sea interface is saturated with water while the air just above the interface typically has relative humidity less than 100%. Thus, moisture tends to evaporate from the ocean and in doing so, the ocean loses heat at a rate of:

$$Q_{lat} = -L(\rho_{fw}E) \qquad [2]$$

where Q_{lat} is the latent heat flux, L is the latent heat of evaporation, ρ_{fw} is the freshwater density, and E is the rate of evaporation. Q_{lat} has units W m^{-2}, and ($\rho_{fw} E$) has units kg s^{-1} m^{-2}. This evaporated moisture can then condense in the atmosphere to form clouds, releasing heat to the atmosphere and affecting the large-scale wind patterns.

0008 Because sensible and latent heat loss are turbulent processes, they also depend on wind speed relative to the ocean surface, S. Using similarity arguments, the latent (Q_{lat}) and sensible (Q_{sen}) heat fluxes can be expressed in terms of 'bulk' properties at and near the ocean surface:

$$Q_{lat} = \rho_a L \, C_E S(q_a - q_s) \qquad [3]$$

$$Q_{sen} = \rho_a c_{pa} C_H \, S(T_a - T_s) \qquad [4]$$

where ρ_a is the air density, c_{pa} is the specific heat of air, C_E and C_H are the transfer coefficients of latent and sensible heat flux, q_s is the saturated specific humidity at T_s, the sea surface temperature, and q_a and T_a are, respectively, the specific humidity and temperature of the air at a few meters above the air–sea interface. The sign convention used here is that a negative flux tends to cool the ocean surface. The transfer coefficients, C_E and C_H, depend upon the wind speed and stability properties of the atmospheric boundary layer, making estimations of the heat fluxes quite difficult. Most algorithms estimate the turbulent heat fluxes iteratively, using first estimates of the heat fluxes to compute the transfer coefficients. Further, the dependence of heat flux on wind speed and SST causes the system to be coupled since the heat fluxes can change the wind speed and SST.

0009 **Figure 2** shows the climatological net surface heat flux, Q_0, and SST for the entire globe. Several patterns are evident. (Note that the spatial structure of the climatological latent heat flux can be inferred
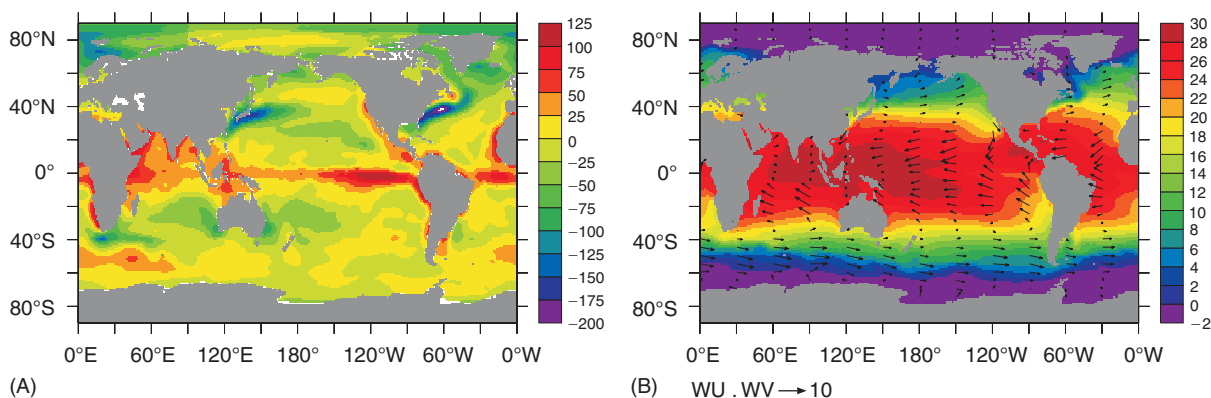
a0157f0002 **Figure 2** Mean climatologies of (A) net surface heat flux (Wm^{-2}) and, (B) sea surface temperature (°C), and surface winds (m s^{-1}). Climatologies provided by da Silva *et al.* (1994). A positive net surface heat flux acts to warm the ocean.

from the climatological evaporation shown in **Figure 3A**.) In general, the tropics are heated more than the poles causing warmer SST in the tropics and cooler SST at the poles. Also, there are significant zonal asymmetries in both the net surface heat flux and SST. The largest ocean surface heat losses occur over the western boundary currents. In these regions, latent and sensible heat loss are enhanced due to the strong winds which are cool and dry as they blow off the continent and over the warm water carried poleward by the western boundary currents. In contrast, the ocean's latent and sensible heat loss are reduced in the eastern boundary region where marine winds blow over the cool water. Consequently, the eastern boundary is a region where the ocean gains heat from the atmosphere. These spatial patterns exemplify the rich variability in the ocean–atmosphere climate system that occurs on a variety of spatial and temporal scales. In particular, seasonal conditions can often be quite different from mean climatology. The seasonal warming and cooling in the north Pacific and eastern equatorial Pacific are discussed later.

Thermal and Haline Buoyancy Fluxes

Since the density of sea water depends on temperature and salinity, air–sea heat and moisture fluxes can change the surface density making the water column more or less buoyant. Specifically, the net surface heat flux (Q_0), rate of evaporation (E) and precipitation (P) can be expressed as a buoyancy flux (B_0), as:

$$B_0 = -g\alpha Q_0/(\rho c_p) + g\beta(E - P)S_0 \qquad [5]$$

where g is gravity, ρ is ocean density, c_p is specific heat of water, S_0 is surface salinity, α is the effective thermal expansion coefficient ($-\rho^{-1}\partial\rho/\partial T$), and β is the effective haline contraction coefficient ($\rho^{-1}\partial\rho/\partial S$). Q_0 has units W m^{-2}, E and P have units m s^{-1}, and B_0 has units m^2 s^{-3}. A negative (i.e., downward) buoyancy flux, due to either surface warming or precipitation, tends to make the ocean surface more buoyant. Conversely, a positive buoyancy flux, due to either surface cooling or evaporation, tends to make the ocean surface less buoyant. As the water column loses buoyancy, it can become
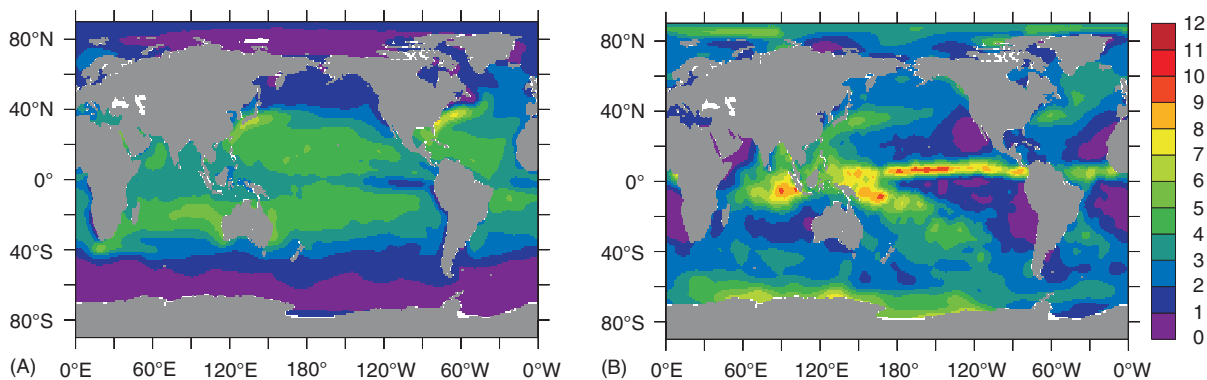

a0157f0003 **Figure 3** Mean climatologies of (A) evaporation, and (B) precipitation from da Silva *et al.* (1994). Both have units mm day^{-1} and share the scale shown on the right.

convectively unstable with heavy water lying over lighter water. Turbulence, generated by the ensuing convective overturning, can then cause deeper, generally cooler, water to be entrained and mixed into the surface mixed layer (**Figure 1**). Thus entrainment mixing typically causes the SST to cool and the mixed layer to deepen. As discussed in the next section, entrainment mixing can also be generated by wind forcing.

Figure 3 shows the climatological evaporation and precipitation fields. Note that in terms of buoyancy, a 20 Wm^{-2} heat flux is approximately equivalent to a 5 mm day^{-1} rain rate. Thus, in some regions of the world oceans the freshwater flux term in eqn [5] dominates the buoyancy flux, and hence is a major factor in the mixed layer thermodynamics. For example, in the tropical regions, heavy precipitation can result in a surface-trapped freshwater pool that forms a shallower mixed layer within a deeper nearly isothermal layer. The difference between the shallower mixed layer of uniform density and the deeper isothermal layer is referred to as a salinity-stratified barrier layer. As the name suggests, a barrier layer can effectively limit turbulent mixing of heat between the ocean surface and the deeper thermocline since entrainment mixing of the barrier layer water into the mixed layer results in no heat flux, if the barrier layer water is the same temperature as the mixed layer. In regions with particularly strong freshwater stratification, solar radiation can warm the waters below the mixed layer, causing the temperature stratification to be inverted (cool water above warmer water). In this case, entrainment mixing produces warming rather than cooling within the mixed layer.

In subpolar latitudes, freshwater fluxes can also dominate the surface layer buoyancy profile. During the winter season, atmospheric cooling of the ocean, and stronger wind mixing leaves the water column isothermal to great depths. Then, wintertime ice formation extracts freshwater from the surface layer, leaving a saltier brine that further increases the surface density, decreases the buoyancy, and enhances the deep convection. This process can lead to deep water formation as the cold and salty dense water sinks and spreads horizontally, forcing the deep, slow thermohaline circulation. Conversely, in summer when the ice shelf and icebergs melt, fresh water is released, and the density in the surface layer is reduced so that the resultant stable halocline (pycnocline) inhibits the sinking of water.

Wind Forcing

The influence of the winds on the ocean circulation and mass field cannot be overstated. Wind blowing over the ocean surface causes a tangential stress ('wind stress') at the interface which acts as a vertical flux of horizontal momentum. Similar to the air–sea fluxes of heat and moisture, this air–sea flux of horizontal momentum, τ_0, can be expressed in terms of bulk properties as:

$$\tau_0 = \rho_a C_D S(u_a - u_s) \qquad [6]$$

where ρ_a is the air density, C_D is the drag coefficient, S is the wind speed relative to the ocean, and $(u_a - u_s)$ is the surface wind relative to the ocean surface flow. The units of the surface wind stress are N m^{-2}. If the upper ocean has a neutral or stable stratification, then the force imparted by the wind stress will generate a momentum flux that is maximum at the surface and decreases with depth through the water column. Consequently, a vertical shear in the velocity develops in the upper ocean mixed layer. The mechanisms by which the momentum flux extends below the interface are not well understood. Some of the wind stress goes into generating ocean surface waves. However, most of the wave momentum later becomes available for generating currents through wave breaking, and wave–wave and wave–current interactions. For example, wave–current interactions associated with Langmuir circulation can set up large coherent vortices which carry momentum to near the base of the mixed layer. As with convective overturning, wind stirring can entrain cooler thermocline water into the mixed layer, producing a colder and deeper mixed layer. Likewise, current shear generated by the winds can cause 'Kelvin Helmholtz' shear instabilities that further mix properties within and at the base of the mixed layer.

Over timescales longer than roughly a day, the Earth's spinning tends to cause a rotation of the vertical flux of momentum. From the non-inertial perspective of an observer on the rotating Earth, the tendency to rotate appears as a force, referred to as the Coriolis force. Consequently, the wind-forced surface layer transport ('Ekman transport') tends to be perpendicular to the wind stress. Because the projection of the Earth's axis onto the local vertical axis (direction in which gravity acts) changes sign at the equator, the Ekman transport is to the right of the wind stress in the Northern Hemisphere and to the left of the wind stress in the Southern Hemisphere. Convergence and divergence of this Ekman transport leads to vertical motion which can deform the thermocline and set the subsurface waters in motion. In this way, meridional variations in the prevailing zonal wind stress drive the steady, large-scale ocean gyres.

The influence of Ekman upwelling on SST can be seen along the eastern boundary of the ocean basins and along the equator (**Figure 2B**). Equatorward winds along the eastern boundaries of the Pacific and Atlantic Oceans cause an offshore-directed Ekman transport. Mass conservation requires that this water be replaced with upwelled water, water that is generally cooler than the surface waters outside the upwelling zone. Likewise, in the tropics, prevailing easterly trade winds cause poleward Ekman transport. At the equator, this poleward flow results in substantial surface divergence and upwelling. As with the eastern boundary, equatorial upwelling results in relatively cold SSTs (**Figure 2B**). Because of the geometry of the continents, the thermal equator favors the Northern Hemisphere and is generally found several degrees of latitude north of the equator.

In the tropics, winds tend to flow from cool SSTs to warm SSTs, where deep atmospheric convection can occur. Thus, surface wind convergence in the inter-tropical convergence zone (ITCZ) is associated with the thermal equator, north of the equator. The relationship between the SST gradient and winds accounts for an important coupling mechanism in the tropics.

The Seasonal Cycle

The North Pacific: A One-dimensional Ocean Response to Wind and Buoyancy Forcing

From 1949 through 1981, a ship (*Ocean Station PAPA*) was stationed in the North Pacific at 50°N 145°W with the primary mission of taking routine ocean and atmosphere measurements. The seasonal climatology observed at this site (**Figure 4**) illustrates a classic near-one-dimensional ocean response to wind and buoyancy forcing. A one-dimensional response implies that only the vertical structure of the ocean is changed by the forcing.

During springtime, layers of warmer and lighter water are formed in the upper surface in response to the increasing solar warming. By summer, this heating has built a stable (buoyant), shallow seasonal pycnocline that traps the warm surface waters. In fall, storms are more frequent and net cooling sets in. By winter, the surface layer is mixed by wind stirring and convective overturning. The summer pycnocline is eroded and the mixed layer deepens to the top of the permanent pycnocline.

To first approximation, horizontal advection does not seem to be important in the seasonal heat
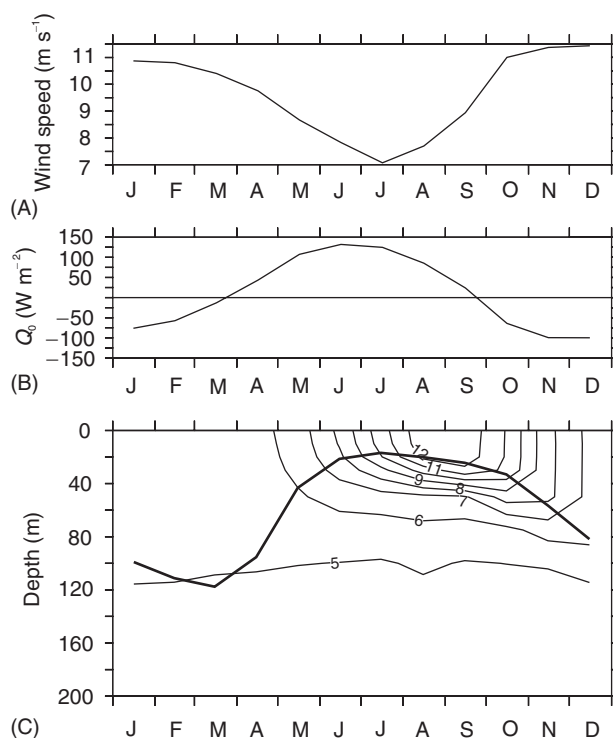
Figure 4 Seasonal climatologies at the ocean weather station PAPA in the north Pacific. (A) Wind speed, (B) net surface heat flux, and (C) upper ocean temperature. The bold line represents the base of the ocean mixed layer defined as the depth where the temperature is 0.5°C cooler than the surface temperature. Wind speed and net surface heat flux climatologies are from da Silva *et al.* (1994).

budget. The progression appears to be consistent with a surface heat budget described by:

$$\partial T/\partial t = Q_0/(\rho c_p H) \qquad [7]$$

where $\partial T/\partial t$ is the local time rate of change of the mixed layer temperature, and H is the mixed layer depth. Since only vertical processes (e.g., turbulent mixing and surface forcing) affect the depth and temperature of the mixed layer, the heat budget can be considered one-dimensional.

A similar one-dimensional progression occurs in response to the diurnal cycle of buoyancy forcing associated with daytime heating and nighttime cooling. Mixed layer depths can vary from just a few meters thick during daytime to several tens of meters thick during nighttime. Daytime and nighttime SST can sometimes differ by over one degree Celsius. However, not all regions of the ocean have such an idealized mixed layer seasonal cycle. Our second example shows a more complicated seasonal cycle in which the tropical atmosphere and ocean are coupled.

### The Eastern Equatorial Pacific: Coupled Ocean–Atmosphere Variability

0021   Because there is no Coriolis turning at the equator, water and air flow are particularly susceptible to horizontal convergence and divergence. Small changes in the winds patterns can cause large variations in oceanic upwelling, resulting in significant changes in SST and consequently in the atmospheric heating patterns. This ocean and atmosphere coupling thus causes initial changes to the system to perpetuate further changes.

0022   At the equator, the sun is overhead twice per year: in March and again in September. Therefore one might expect a semiannual cycle in the mixed layer properties. Although this is indeed found in some parts of the equatorial oceans (e.g., in the western equatorial Pacific), in the eastern equatorial Pacific the annual cycle dominates. During the warm season (February–April), the solar equinox causes a maximum in insolation, equatorial SST is warm and the meridional SST gradient is weak.

Consequently, the ITCZ is near the equator, and often a double ITCZ is observed that is symmetric about the equator. The weak winds associated with the ITCZ cause a reduction in latent heat loss, wind stirring, and upwelling, all of which lead to further warming of the equatorial SSTs. Thus the warm SST and surface heating are mutually reinforcing.

0023   Beginning in about April–May, SSTs begin to cool in the far eastern equatorial Pacific, perhaps in response to southerly winds associated with the continental monsoon. The cooler SSTs on the equator cause an increased meridional SST gradient that intensifies the southerly winds and the SST cooling in the far eastern Pacific. As the meridional SST gradient increases, the ITCZ begins to migrate northward. Likewise, the cool SST anomaly in the far east sets up a zonal SST gradient along the equator that intensifies the zonal trade winds to the west of the cool anomaly. These enhanced trade winds then produce SST cooling (through increased upwelling, wind stirring and latent heat loss) that spreads westward (**Figure 5**).
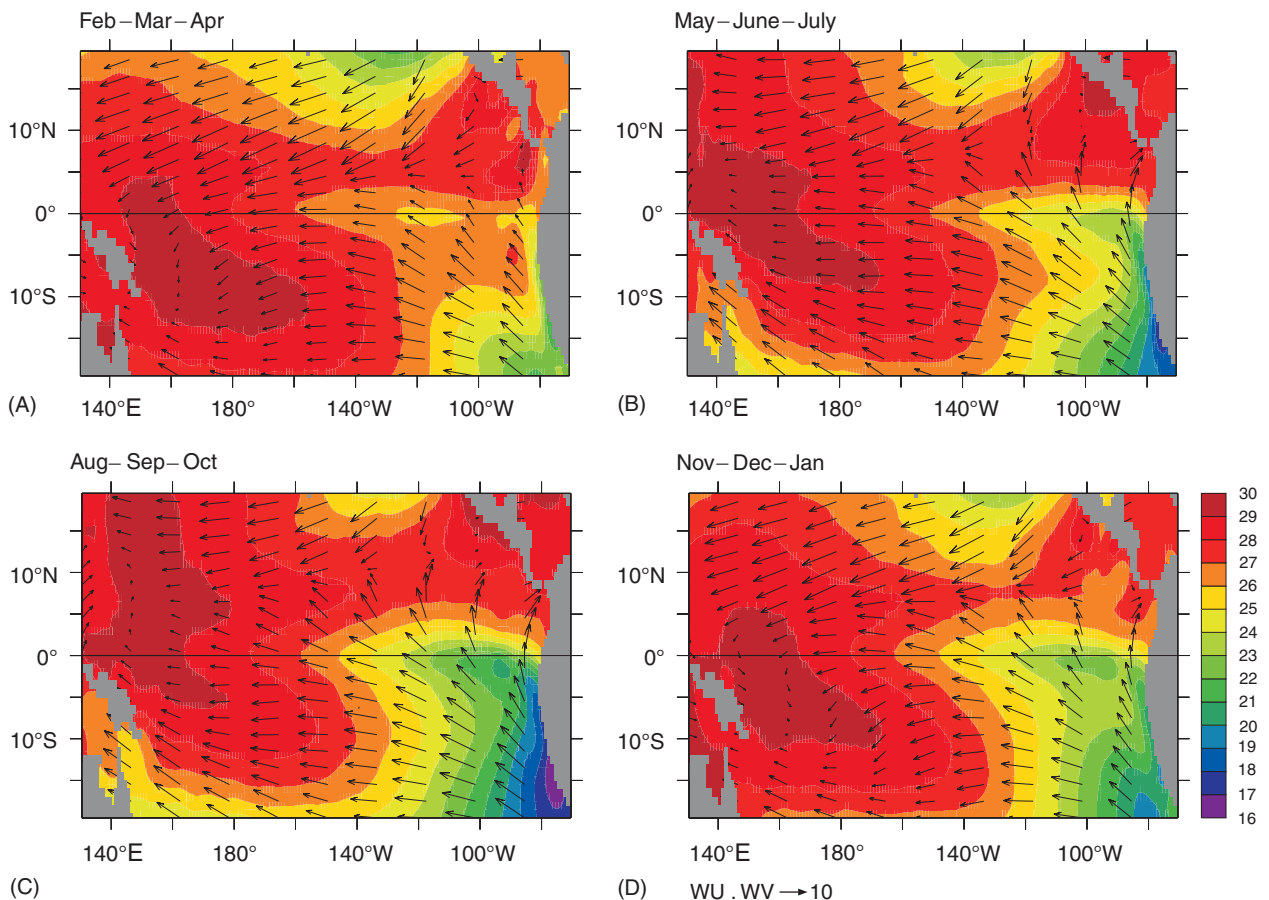
a0157f0005   **Figure 5**   Seasonal climatologies of the tropical Pacific sea surface temperature (°C) and wind (m s⁻¹) for (A) February–March–April, (B) May–June–July, (C) August–September–October, and (D) November–December–January. Winds are from da Silva *et al.* (1994). SSTs are from Reynolds and Smith (1994).

0024 By September, the equatorial cold-tongue is fully formed. Stratus clouds, which tend to form over the very cool SSTs in the tropical Pacific, cause a reduction in solar radiation, despite the equinoctial increase. The large meridional gradient in SST associated with the fully formed cold tongue causes the ITCZ to be at its northernmost latitude. After the cold tongue is fully formed, the reduced zonal SST gradient within the cold tongue causes the trade winds to weaken there, leading to reduced SST cooling along the equator. Finally, by February, the increased solar radiation associated with the approaching vernal equinox causes the equatorial SSTs to warm and the cold tongue to disappear, bringing the coupled system back to the warm season conditions.

Conclusion

0025 Because the ocean mixed layer responds so rapidly to surface-generated turbulence through wind and buoyancy forced processes, the surface mixed layer can often be modeled successfully using one-dimensional (vertical processes only) physics. Surface heating and cooling cause the ocean surface to warm and cool; evaporation and precipitation cause the ocean surface to become saltier and fresher. Stabilizing buoyancy forcing, whether from net surface heating or precipitation, stratifies the surface and isolates it from the deeper waters. Whereas wind stirring and destabilizing buoyancy forcing generate surface turbulence that cause the surface properties to mix with deeper water. Eventually, however, one-dimensional models drift away from observations, particularly in regions with strong ocean–atmosphere coupling and oceanic current structures. The effects of horizontal advection are explicitly not included in one-dimensional models. Likewise, vertical advection depends on horizontal convergences and divergences and therefore is not truly a one-dimensional process. Finally, wind and buoyancy forcing can themselves depend on the horizontal SST patterns, blurring the distinction between forcing and response.

0026 Although the mixed layer is the principal region of wind and buoyancy forcing, ultimately the effects are felt throughout the world's oceans. Both the wind-driven motion below the mixed layer and the thermohaline motion in the relatively more quiescent deeper ocean originate through forcing in the surface layer that causes an adjustment in the mass field (i.e., density profile). In addition, buoyancy and wind-forcing in the upper ocean define the property characteristics for all the individual major water masses found in the world oceans. On a global scale, there is surprisingly little mixing between water masses once they acquire the characteristic properties at their formation region and are vertically subducted or convected from the active surface layer. As these subducted water masses circulate through the global oceans and later outcrop, they can contain the memory of their origins at the surface through their water mass properties and thus can potentially induce decadal and centennial modes of variability in the ocean–atmosphere climate system.

See also

Breaking Waves and Near-surface Turbulence. Ekman Transport and Pumping. Langmuir Circulation and Instability. Pacific Ocean Equatorial Currents. Penetrating Shortwave Radiation. Thermohaline Circulation. Upper Ocean Heat and Freshwater Budgets. Upper Ocean Responses to Strong Forcing Events. Upper Ocean Time and Space Variability. Upper Ocean Vertical Structure. Wind Driven Circulation. Water Types and Water Masses.

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