14

Upwelling

Summary: Coastal upwelling is a phenomenon implicating wind forcing, Ekman dynamics, geostrophy, and stratification. It also has important biological consequences. Two other types of upwelling, along the equator and along the marginal ice zone, are briefly described.

14-1 THE UPWELLING PROCESS

Winds blowing over the ocean generate Ekman layers and currents. The depth-averaged currents, called the Ekman drift, forms an angle with the wind, which was found to be 90° (to the right in the Northern Hemisphere) according to a simple theory (Section 5-4). So, when a wind blows along a coast, it generates an Ekman drift directed either onshore or offshore, to which the coast stands as an obstacle. The drift is offshore if the wind blows with the coast on its left (right) in the Northern (Southern) Hemisphere (Figure 14-1). If this is the case, water depletion occurs in the upper layers, and a low pressure sets in, forcing waters from below to move upward and replenish the space vacated by the offshore drift. This phenomenon is called *coastal upwelling*. The upward movement calls for a replenishment at the lower levels, which is accomplished by an onshore flow. To recapitulate, a wind blowing along the coast (with the coast on the

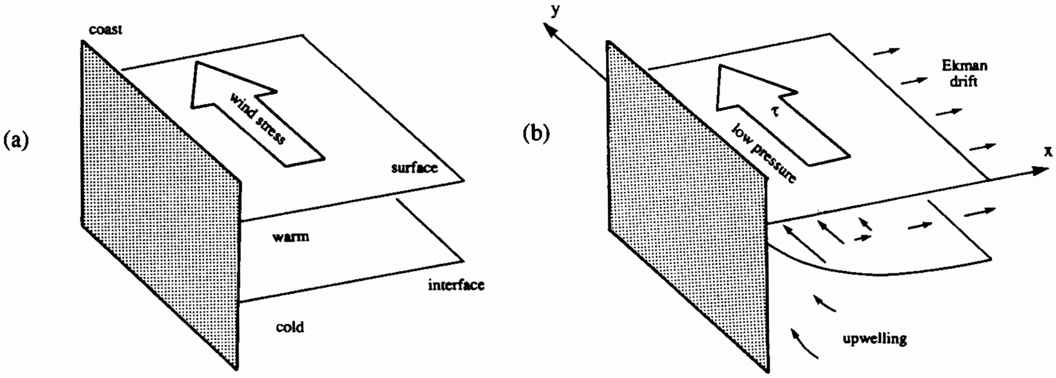

Figure 14-1 Schematic development of coastal upwelling.

left or the right in, respectively, the Northern or the Southern Hemisphere) sets an offshore current in the upper levels, an upwelling at the coast, and an onshore current at lower levels. This circulation in the cross-shore vertical plane is not the whole story, however. The low pressure created along the coast also sustains, via geostrophy, a longshore current, while vertical stretching in the lower layer generates relative vorticity and a shear flow. Thus, the flow pattern is rather complex.

At the root of coastal upwelling is a divergent Ekman drift. And, we can easily conceive of other causes besides a coastal boundary for such divergence. Two other upwelling situations are noteworthy: one along the equator and the other at high latitudes.

Along the equator, the trade winds blow almost steadily from east to west. On the northern side of the equator, the Ekman drift is to the right, or away from the equator, but on the southern side, it is to the left, again away from the equator (Figure 14-2a). Consequently, horizontal divergence occurs along the equator, and mass conservation requires upwelling (Yoshida, 1959; Gill, 1982, Chapter 11).

At high latitudes, upwelling frequently occurs along the ice edge, in the so-called marginal ice zone. A uniform wind exerts different stresses on ice and open water; in its turn, the moving ice exerts a stress on the ocean beneath. The net effect is a complex distribution of stresses and velocities at various angles, with the likely result that the ocean currents at the ice edge do not match (Figure 14-2b). For certain angles between wind and ice edge, these currents diverge, and upwelling again takes place to compensate for the divergence of the horizontal flow (Häkkinen, 1990).

The upwelling phenomenon, especially the coastal type, has been the subject of considerable attention, chiefly because of its relation to biological oceanography and, from there, to fisheries. In brief, small organisms in the ocean (phytoplankton) proliferate when two conditions are met: sunlight and a supply of nutrients. In general, nutrients lie in the deeper waters, below the reach of sunlight, and so the waters tend to lack either nutrients or sunlight. The major exceptions are the upwelling regions, where deep, nutrient-rich waters rise to the surface, receive sunlight, and stimulate biological activity. Upwelling-favorable winds most generally occur along the west

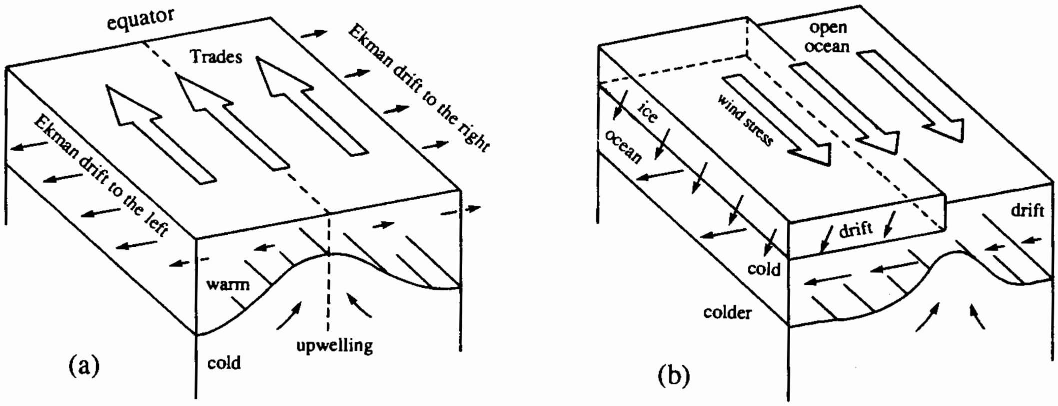

Figure 14-2 Other types of oceanic upwelling: (a) equatorial upwelling, (b) upwelling along the ice edge.

coasts of continents where the prevailing winds blow toward the equator. For a review of observations and a discussion of the biological implications of coastal upwelling, the interested reader is referred to the volume edited by Richards (1981).

14-2 A SIMPLE MODEL OF COASTAL UPWELLING

Consider a reduced-gravity ocean on an f-plane ($f > 0$), bounded by a vertical wall and subjected to a surface stress acting with the wall on its left (Figure 14-1a). The upper moving layer, meant to include the entire vertical extent of the Ekman layer, supports an offshore drift current. The lower layer is, by virtue of the choice of a reduced-gravity model, infinitely deep and motionless. In the absence of longshore variations, the equations of motion are

$$\frac{\partial u}{\partial t} + u \frac{\partial u}{\partial x} - fv = -g' \frac{\partial h}{\partial x} \tag{14-1}$$

$$\frac{\partial v}{\partial t} + u \frac{\partial v}{\partial x} + fu = \frac{\tau}{\rho_0 h} \tag{14-2}$$

$$\frac{\partial h}{\partial t} + \frac{\partial}{\partial x}(hu) = 0, \tag{14-3}$$

where x is the offshore coordinate, τ is the longshore wind stress, and all other symbols are conventional (Figure 14-1b).

Despite its apparent simplicity, the preceding set of equations is nonlinear, and no analytical solution is known. We therefore linearize these equations by assuming that the wind stress τ and, in turn, the ocean's reaction are weak. Noting $h = H - \eta$, where H is the depth of the undisturbed upper layer and η is the small upward displacement of the interface, we write

$$\frac{\partial u}{\partial t} - fv = g' \frac{\partial \eta}{\partial x} \tag{14-4}$$

$$\frac{\partial v}{\partial t} + fu = \frac{\tau}{\rho_0 H} \tag{14-5}$$

$$-\frac{\partial \eta}{\partial t} + H \frac{\partial u}{\partial x} = 0. \tag{14-6}$$

This set of equations contains two independent x-derivatives and calls for two boundary conditions: u vanishes at the coast ($x = 0$) and η vanishes far offshore ($x \to +\infty$).

The solution to the problem depends on the initial conditions, which may be taken to correspond the state of the rest ($u = v = \eta = 0$). Yoshida (1955) is credited with the first derivation of this solution (extended to two moving layers). However, because of the fluctuating nature of winds, upwelling is rarely an isolated event in time, and we prefer to investigate the periodic solutions to the preceding linear set of equations. Taking $\tau = \tau_0 \sin \omega t$, where τ_0 is a constant, we note that the solution must be of the type $u = u_0(x) \sin \omega t$, $v = v_0(x) \cos \omega t$ and $\eta = \eta_0(x) \cos \omega t$. Substitution and solution of the remaining ordinary differential equations in x yield

$$u = \frac{f\tau_0}{\rho_0 H\,(f^2 - \omega^2)} \left[1 - \exp\left(-\frac{x}{R_\omega}\right)\right] \sin \omega t \tag{14-7}$$

$$v = \frac{\omega\tau_0}{\rho_0 H\,(f^2 - \omega^2)} \left[1 - \frac{f^2}{\omega^2} \exp\left(-\frac{x}{R_\omega}\right)\right] \cos \omega t \tag{14-8}$$

$$\eta = \frac{-fR_\omega \tau_0}{\rho_0 g' H\omega} \exp\left(-\frac{x}{R_\omega}\right) \cos \omega t, \tag{14-9}$$

where R_ω is a modified deformation radius defined as

$$R_\omega = \sqrt{\frac{g'H}{f^2 - \omega^2}}\,. \tag{14-10}$$

From the preceding solution, we conclude that the upwelling or downwelling signal is trapped along the coast within a distance on the order of R_ω. Far offshore ($x \to \infty$), the interfacial displacement vanishes, and the flow field reduces to the Ekman drift

$$u_{Ek} = \frac{\tau_0}{\rho_0 f H} \sin \omega t, \quad v_{Ek} = 0, \tag{14-11}$$

on which are superimposed inertial oscillations. At long periods such as weeks and months ($\omega \ll f$), the distance R_ω becomes the radius of deformation, the vertical interfacial displacements become very large (indeed, the wind blows more steadily in one direction before it reverses), and the far-field inertial oscillations become much smaller than the Ekman drift.

At superinertial frequencies ($\omega > f$), the quantity R_ω becomes imaginary, indicating that the solution does not decay away from the coast but instead oscillates.

Physically, the ocean's response is not trapped near the coast and inertia-gravity waves (Section 6-3) are excited. These radiate outward, filling the entire basin.

14-3 FINITE-AMPLITUDE UPWELLING

If the wind is sufficiently strong or is blowing for a sufficiently long time, the density interface can rise to the surface, forming a front. Continued wind action displaces this front offshore and exposes the colder waters to the surface. This mature state is called *full upwelling* (Csanady, 1977). Obviously, the previous linear theory is not applicable in such case.

Because of the added complications arising from the nonlinearities, let us now restrict our investigation to the final state of the ocean after a wind event of finite duration. Equation (14-2), expressed as

$$\frac{d}{dt}(v + fx) = \frac{\tau}{\rho_0 h}, \tag{14-12}$$

where $d/dt = \partial/\partial t + u\,\partial/\partial x$ is the time derivative following a fluid particle in the offshore direction, can be integrated over time to yield:

$$(v + fx)_{\text{at end of event}} - (v + fx)_{\text{initially}} = I. \tag{14-13}$$

The *wind impulse* I is the integration of the wind-stress term, $\tau/\rho_0 h$, over time and following a fluid particle. Although the wind impulse received by every parcel cannot be precisely determined, it can be estimated by assuming that the wind event is relatively brief. The time integral can then be approximated by using the local stress value and replacing h by H:

$$I \simeq \frac{1}{\rho_0 H}\int_{\text{event}} \tau \, dt. \tag{14-14}$$

If the initial state is one of rest, relation (14-13) implies that a particle initially at distance X from the coast is at distance x immediately after the wind ceases and has a longshore velocity v such that

$$v + fx - fX = I. \tag{14-15}$$

During the oceanic adjustment that follows until equilibrium is reached, equation (14-12) (with $\tau = 0$) implies that the quantity $v + fx$ is a conserved quantity, and relation (14-15) continues to hold beyond the time when the wind ceased.

If the wind is laterally uniform while it blows, no vorticity is imparted to fluid parcels, and potential vorticity is conserved during the wind event as well as during the following adjustment:

$$\frac{1}{h}\left(f + \frac{\partial v}{\partial x}\right) = \frac{f}{H}. \tag{14-16}$$

Once a steady state has been achieved, there is no longer any offshore velocity ($u = 0$), according to (14-3). The remaining equation, (14-1), reduces to a simple geostrophic balance, which together with (14-16) provides the solution:

$$h = H - A \exp\left(-\frac{x}{R}\right) \tag{14-17}$$

$$v = A \sqrt{\frac{g'}{H}} \exp\left(-\frac{x}{R}\right), \tag{14-18}$$

where R is now the conventional radius of deformation $(g'H)^{1/2}/f$. The constant of integration A represents the amplitude of the upwelled state and is related to the wind impulse via (14-15). Two possible outcomes must be investigated: Either the interface has not risen to the surface (Figure 14-3, case I) or it has outcropped, forming a front and leaving cold waters exposed to the surface near the coast (Figure 14-3, case II).

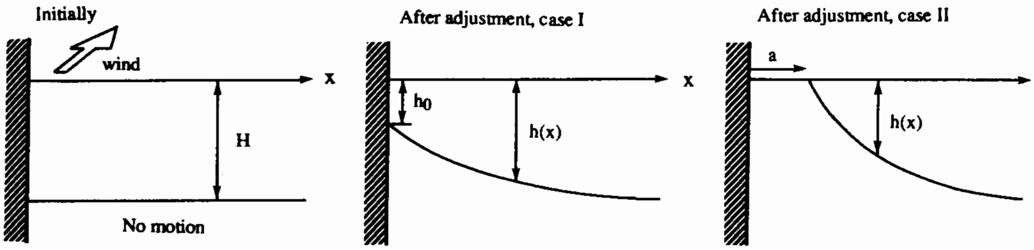

Figure 14-3 The two possible outcomes of coastal upwelling after a longshore wind of finite duration. After a weak or brief wind event (case I), the interface has upwelled but not to the point of reaching the surface. A strong or prolonged wind event (case II) causes the interface to reach the surface, where it forms a front; this front is displaced offshore, leaving cold waters from below exposed to the surface. This latter case corresponds to a mature upwelling that favors biological activity.

In case I, the particle initially against the coast ($X = 0$) is still there ($x = 0$), and relation (14-15) yields $v(x = 0) = I$. Solution (14-18) meets this condition if $A = I(H/g')^{1/2}$. The depth along the coast, $h(x = 0) = H - A$, must be positive requiring $A \leq H$; that is, $I \leq (g'H)^{1/2}$. In other words, the no-front situation or partial upwelling of case I occurs if the wind is sufficiently weak or sufficiently brief that its resulting impulse is less than the critical value $(g'H)^{1/2}$.

In the more interesting case II, the front has been formed, and the particle initially against the coast ($X = 0$) is now at some offshore distance ($x = a \geq 0$), marking the position of the front. There the layer depth vanishes, $h(x = a) = 0$, and solution (14-17) yields $A = H \exp(a/R)$. The longshore velocity at the front is, according to (14-18), $v(x = a) = (g'H)^{1/2}$. Finally, relation (14-15) leads to the determination of the offshore displacement a in terms of the wind impulse:

$$a = \frac{I}{f} - R. \tag{14-19}$$

Since this displacement must be a positive quantity, it is required that $I \geq (g'H)^{1/2}$. Physically, if the wind is sufficiently strong or sufficiently prolonged, so that the net impulse is greater than the critical value $(g'H)^{1/2}$, the density interface rises to the surface and forms a front that migrates away from shore, leaving cold waters from below exposed to the surface. (Note how the conditions for the realizations of cases I and II complement each other.)

Formula (14-19) has a simple physical interpretation, as sketched in Figure 14-4. The offshore Ekman velocity u_{Ek} is the velocity necessary for the Coriolis force to balance the longshore wind stress:

$$u_{Ek} = \frac{\tau}{\rho_0 f h},$$

according to (5-19a). Integrated over time, this yields a net offshore displacement proportional to the wind impulse:

$$x_{Ek} = \frac{I}{f}.$$

If we were now to assume that the wind is responsible for an offshore shift of this magnitude, while the surface waters are moving as a solid slab, we would get the intermediate structure of Figure 14-4. But such a situation cannot persist, and an adjustment must follow, causing an onshore spread similar to that considered in Section 13-2—that is, over a distance equal to the deformation radius; hence we have the final structure of Figure 14-4 and formula (14-19).

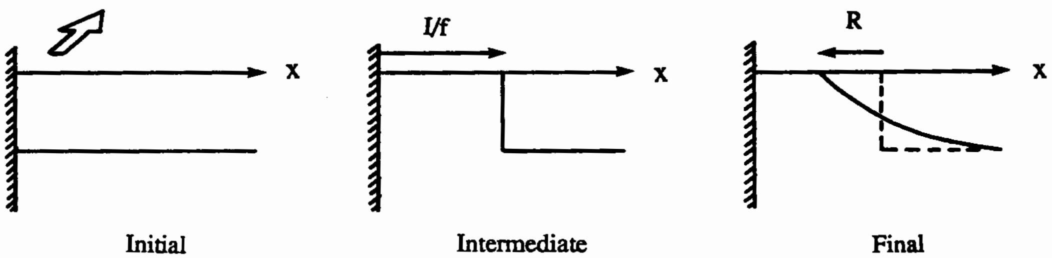

Figure 14-4 Sketch decomposing the formation of a coastal-upwelling front as a two-stage process: first, an offshore displacement in response to the wind, followed by a geostrophic adjustment.

14-4 VARIABILITY OF THE UPWELLING FRONT

Up to this point, we have considered only processes operating in the offshore direction or, equivalently, an upwelling that occurs uniformly along an entire coast. In reality, wind events are usually localized, the coastline is not straight, and upwelling is not at all uniform. A local upwelling sends a wave signal along the coast, taking the form of an internal Kelvin wave, which in the Northern Hemisphere propagates with the coast on its right. This redistribution of information not only decreases the rate of upwelling in the forced region but also generates upwelling in other, unforced areas. As a result, models of upwelling must retain a sizable portion of the coast and both spatial and temporal variations of the wind field (Crépon and Richez, 1982; Brink, 1983).

Adding to this variability are intrinsic instabilities of the upwelling front, because the front is a region of highly sheared currents. In the two-layer formulation presented in the previous section, this shear is manifested by a discontinuity of the current at the front. The warm layer develops anticyclonic vorticity (i.e., counter to the rotation of the earth) under the influence of vertical squeezing and flows alongshore in the direction of the wind. On the other side of the front, the exposed lower layer is stretched vertically, develops cyclonic vorticity (i.e., in the same direction as the rotation of the earth) and flows upwind. The currents on each side of the front thus flow in opposite directions, causing a large shear, which, as we have seen (Chapter 7), is vulnerable to instabilities. In addition to the kinetic-energy supply in the horizontal shear (barotropic instability), potential energy can also be released from the stratification by a spreading of the warm layer (baroclinic instability; see Chapter 16). The situation, undoubtedly complex, is not entirely understood at the present time. One thing is certain, however: Irregularities in the topography and the shape of the coastline play influential roles. Offshore jets of cold, upwelled waters have been observed to form near capes; these jets cut through the front, forge their way through the warm layer, and eventually split to form a pair of counterrotating vortices (Flament et al., 1985). The net effect is the persistence of vigorous, large-scale turbulence along upwelling fronts (Figure 14-5).

PROBLEMS

14-1. Using the linearized equations for a two-layer ocean (undisturbed depths H_1 and H_2) over a flat bottom and subject to a spatially uniform wind stress,

$$\frac{\partial u_1}{\partial t} - f v_1 = g' \frac{\partial \eta}{\partial x} - \frac{1}{\rho_0}\frac{\partial p_2}{\partial x}, \quad \frac{\partial u_2}{\partial t} - f v_2 = -\frac{1}{\rho_0}\frac{\partial p_2}{\partial x},$$

$$\frac{\partial v_1}{\partial t} + f u_1 = \frac{\tau}{\rho_0 H_1}, \quad \frac{\partial v_2}{\partial t} + f u_2 = 0,$$

$$-\frac{\partial \eta}{\partial t} + H_1 \frac{\partial u_1}{\partial x} = 0, \quad \frac{\partial \eta}{\partial t} + H_2 \frac{\partial u_2}{\partial x} = 0,$$

study the upwelling response to an oscillating wind stress. The boundary conditions are: no flow at the coast ($u_1 = u_2 = 0$ at $x = 0$), no vertical displacements, and no pressure anomaly at large distances ($\eta \to 0$ and $p_2 \to 0$ as $x \to +\infty$). Discuss how the dynamics of the upper layer are affected by the presence of an active lower layer and what happens in the lower layer.

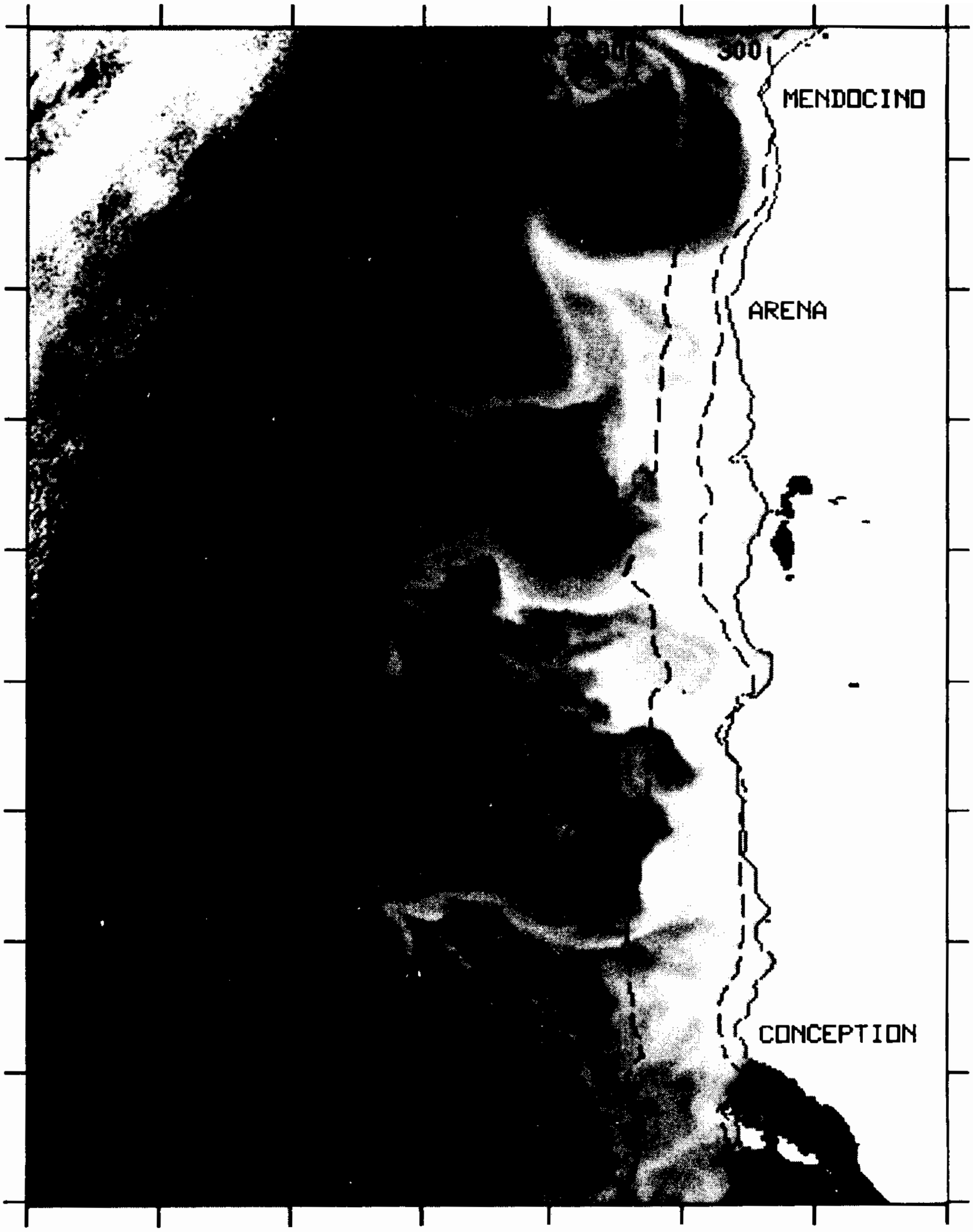

Figure 14-5 Satellite infrared image of the Pacific Ocean off the coast of California, taken on 15 June 1981. The spacing of the tic marks is 100 km. Light gray shades indicate cold water. Cold upwelled water is seen on the continental shelf and slope shown by the 300-m and 3000-m isobaths (dashed lines). Note the convoluted form of the upwelling front and the offshore jets between Point Conception (120° 30′W, 34° 30′N) and Cape Mendocino (124° 20′W, 40° 20′N). (From P. Flament et al., *Journal of Geophysical Research—Oceans*, 90, 11765–78, 1985, copyright by the American Geophysical Union.)

14-2. Demonstrate the assertion made in the text (Section 14-3) that the potential vorticity is conserved if the wind stress is spatially uniform.

14-3. A coastal ocean at midlatitude ($f = 10^{-4}\ s^{-1}$) has a 50-m thick warm layer capping a much deeper cold layer. The relative density difference between the two layers is $\Delta\rho/\rho_0 = 0.002$. A uniform wind exerting a surface stress of 0.4 N/m^2 lasts for three days. Show that the resulting upwelling includes outcropping of the density interface. What is the offshore distance of the front?

14-4. Generalize to the two-layer ocean the theory for the steady adjusted state following a wind event of given impulse. For simplicity, consider only the case of equal initial layer depths ($H_1 = H_2$).

14-5. Because of the roughness of the ice, the stress communicated to the water is substantially larger in the presence of sea ice than in the open sea. Assuming that the ice drift is at 20° to the wind, that the water drift is at 90° to the wind (in the open) and to the ice drift (under ice), and that the stress on the water surface is twice as large under ice than in the open sea, determine which wind directions with respect to the ice-edge orientation are favorable to upwelling.

Kozo Yoshida

1922 – 1978

In the early years of his professional career, Kozo Yoshida studied long (tsunami) and short (wind) waves. Later, during a stay at the Scripps Institution of Oceanography, he turned to the investigation of the upwelling phenomenon, which was to become his lifelong interest. His formulations of dynamic theories for both coastal and equatorial upwelling earned him respect and fame. A wind-driven surface eastward current along the equator is called a Yoshida jet. In his later years, he also became interested in the Kuroshio, a major ocean current off the coast of Japan, wrote several books, and promoted oceanography among young scientists. Known to be very sincere and logical, Yoshida did not shun administrative responsibilities and emphasized the importance of international cooperation in postwar Japan. (*Photo courtesy of Toshio Yamagata.*)