Ocean waves

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Propagating oscillations in the ocean which carry energy and momentum from one region to another. Most ocean waves are caused directly or indirectly by wind blowing across the sea surface. Many waves can propagate through the ocean thousands of miles from where they are generated.

Surface waves

Ocean surface waves are propagating disturbances at the atmosphere-ocean interface. They are the most familiar ocean waves. Surface waves are also seen on other bodies of water, including lakes and rivers. *See also:* SURFACE WAVES.

A simple sinusoidal wave train is characterized by three attributes: wave height (H), the vertical distance from trough to crest; wavelength (L), the horizontal crest-to-crest distance; and wave period (T), the time between passage of successive crests past a fixed point. The phase velocity ($C = L/T$) is the speed of propagation of a crest. For a given ocean depth (h), wavelength increases with increasing period. The restoring force for these surface waves is predominantly gravitational. Therefore, they are known as surface gravity waves, unless their wavelength is shorter than 1.8 cm (0.7 in.), in which case surface tension provides the dominant restoring force.

Classification

Surface gravity waves may be classified according to the nature of the forces producing them. Tides are ocean waves induced by the varying gravitational influence of the Moon and Sun. They have long periods, usually 12.42 h for the strongest constituent. Storm surges are individual waves produced by the wind and dropping barometric pressure associated with storms; they characteristically last several hours (**Fig. 1**). Earthquakes or other large, sudden movements of the Earth's crust can cause waves, called tsunamis, which typically have periods of less than an hour. Wakes are waves resulting from relative motion of the water and a solid body, such as the motion of a ship through the sea or the rapid flow of water around a rock. Wind-generated waves, having periods from a fraction of a second to tens of seconds, are called wind waves. Like tides, they are ubiquitous in the ocean and continue to travel well beyond their area of generation. The ocean is never completely calm. *See also:* STORM SURGE; TIDE; TSUNAMI.

Fig. 1 **North Pacific storm waves as seen from the NOAA *M/V Noble Star,* Winter 1989. *(NOAA/Department of Commerce*)**

Wind waves

The growth of wind waves by the transfer of energy from the wind is not fully understood. At wind speeds less than 1.1 m/s (2.5 mi/h), a flat water surface remains unruffled by waves. Once generated, waves gain energy from the wind by wave-coupling of pressure fluctuations in the air just above the waves. For waves traveling slower than the wind, secondary, wave-induced airflows shift the wave-induced pressure disturbance downwind so the lowest pressure is ahead of the crests. This results in energy transfer from the wind to the wave and hence growth of the wave.

If a constant wind blows over a sufficient length of ocean, called the fetch, for a sufficient length of time, a wave field develops whose statistical characteristics depend only on wind velocity. In particular, the spectrum of sea-surface elevation for such a fully developed sea has the form of Eq. (**1**),

$$S(f) = A\frac{g^2}{f^5}e^{-1.25(f_m/f)^4} \tag{1}$$

where f is frequency ($= 1/T$), $g = 9.8$ m/s^2 (32 ft/s^2) is gravitational acceleration, $f_m = 0.13\, g/U$ is the frequency of the spectral peak (U = wind speed at 10 m or 32.8 ft elevation) and $A = 5.2 \times 10^{-6}$ is a constant.

The fetch is limited near a coast with the wind blowing offshore and the waves grow as they propagate toward the open ocean. In such a limited-fetch situation, Eq. (1) is modified: A and f_m become dependent on the fetch length and the peak in the spectrum is enhanced. **Figure 2** shows spectral forms of waves generated by a moderate breeze for various fetches and in the open sea. For faster wind speeds, the spectral peaks grow in height and shift to lower frequencies.

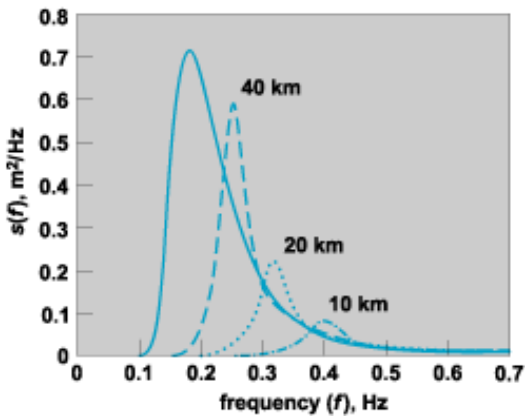

Fig. 2 **Spectra of waves from a 7 m/s (16 mi/h) wind in the open sea (solid line) and for three limited fetches: 40, 20, 10 km (25, 12, 6 mi). (*Theoretical relations are from W. J. Pierson and L. Moskovitz, J. Geophys. Res., 69:5181–5190, 1964, for the open sea and K. Hasselmann et al., Deutsch. Hydrogr. Z., Reihe A (8°), Nr. 12, 1973, for fetch-limited seas*)**

Even when the mean wind blows from a single direction, the surface waves that it generates are seen to travel in a variety of directions centered on the downwind direction. The directional spectrum of such a wave field can be approximately represented by a formula, such as Eq. (**2**),

$$S(f,\theta) = \begin{cases} S(f)\dfrac{2}{\pi}\cos^2\theta & \text{if } \theta \le 90^\circ \\ 0 & \text{if } \theta > 90^\circ \end{cases} \tag{2}$$

where θ is the angle of wave propagation relative to the downwind direction.

An observer asked to estimate average wave height typically gives a value that is about the average height of the highest one-third of the waves actually present. This statistic, represented as $H_{1/3}$, is called the significant wave height. In the open sea, Eq. (**3**)

$$H_{1/3} = 0.24\frac{U^2}{g} \tag{3}$$

applies. If a wind blows steadily over a known fetch for a period of time, the resulting $H_{1/3}$ may be estimated from **Fig. 3**.

Because of viscosity, surface waves lose energy as they propagate, short-period waves being dampened more rapidly than long-period waves. Waves with long periods (typically 10 s or more) can travel thousands of

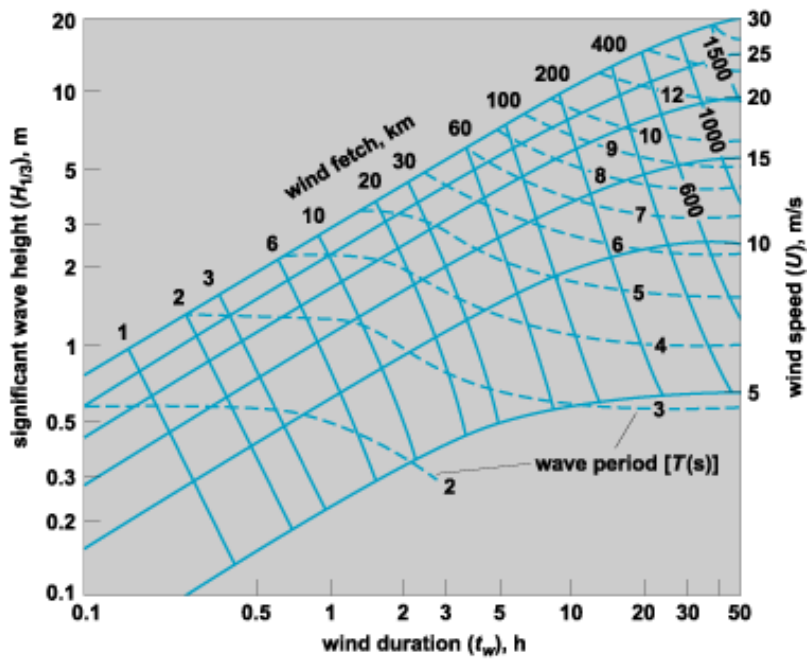

Fig. 3 **Relationship of significant wave height $H_{1/3}$ and characteristic wave period T_m (=1/f_m) to wind speed, duration and fetch. To estimate the wave characteristics resulting from a wind of velocity U which has blown steadily for a time t_w, find one of the six heavy curves labeled on the right with the appropriate U, follow this curve to the correct t_w, and, at that point, read the value of $H_{1/3}$. The position of this point, relative to the broken curves, indicates the wave period T_m. If the waves are fetch-limited, $H_{1/3}$ grows with time only until the heavy curve intersects the appropriate fetch line, at which point $H_{1/3}$ and T_m remain fixed. Open-sea conditions exist on the right side of the diagram. (*After World Meteorological Organization, Guide to Wave Analysis and Forecasting, no. 702, 1988*)**

kilometers with little energy loss. Such waves, generated by distant storms, are called swell. Equations (**1**), (**2**), (**3**) and **Figs. 2** and **3** assume that the waves present were generated by local wind, with no significant swell present. *See also:* VISCOSITY.

The highest wind waves are produced by large intense storm systems that last for a day or longer. Such systems of very low atmospheric pressure form in the Gulf of Alaska, the region around Iceland and the Weddell Sea. Off the west coast of Canada, there have been several measurements of individual waves with heights around 30 m (100 ft). Northwest of Hawaii, on February 7, 1933, the Navy tanker USS *Ramapo* encountered the largest open-ocean wind waves ever reliably observed with heights that were reported to be at least 34 m (112 ft).

When waves propagate into an opposing current, they grow in height. For example, when swell from a Weddell Sea storm propagates northeastward into the southwestward-flowing Agulhas Current off South Africa, high steep waves are formed. Many large ships in this region have been severely damaged by such waves.

Because actual ocean waves consist of many components with different periods, heights and directions, occasionally a large number of these components can, by chance, come in phase with one another, creating a freak wave with a height several times the significant wave height of the surrounding sea. According to linear theory, waves with different periods propagate with different speeds in deep water and hence the wave components remain in phase only briefly. But nonlinear effects are bound to be significant in a large wave. In such a wave, the effects of nonlinearity can compensate for those of dispersion, allowing a solitary wave to propagate almost unchanged. Consequently, a freak wave can have a lifetime of a minute or two. *See also:* SOLITON.

Linear theory

For waves sufficiently small that linear theory applies, Eq. (**4**)

$$C = \begin{cases} (g/k)^{1/2} & \text{deep water} \quad h > 0.4L \\ (gh)^{1/2} & \text{shallow water} \quad h < 0.04L \end{cases} \tag{4}$$

gives the phase velocity to an accuracy of 1%. Note that wavenumber $k = 2\pi/L$. *See also:* WAVE MOTION.

Ocean wave energy E per unit surface area depends only on wave height [Eq. (**5**)].

$$E = \frac{1}{8}\rho g H^2 \tag{5}$$

This energy propagates at the group velocity [Eq. (**6**)]

$$C_g = \begin{cases} \frac{1}{2}C & \text{deep water} \\ C & \text{shallow water} \end{cases} \tag{6}$$

and produces a power flux per unit distance along the wave front [Eq. (**7**)].

$$P = C_g E \tag{7}$$

For example, waves in the deep sea with 1.8-m (6-ft) height and 10-s period carry a power flux of 30 kW/m (13 hp/ft). Available power fluxes of this magnitude are representative of many coastal regions and represent a

substantial renewable energy resource. Various devices with efficiencies of 50% or greater have been developed to convert this wave energy to electrical energy. *See also:* HYDROELECTRIC GENERATOR.

Surface-gravity waves cause pressure fluctuations and particle motions that are largest near the surface and decrease with depth. In deep water, this dependence on depth d below the still-water level is exponential: e^{-kd} .

Stokes drift

To first-order (linear theory), particle orbits resulting from wave motion are closed loops. But since a particle moves in the direction of wave propagation in the loop's upper part and in the reverse direction in its lower part, the forward motion is at a shallower level than the reverse motion. Consequently, the forward motion is slightly stronger than the reverse motion and the orbital loops do not quite close. As a result, there is a second-order (nonlinear) net particle motion called Stokes drift. In deep water, the velocity associated with this motion is in the direction of wave propagation [Eq. (**8**)].

$$\bar{u} = C(ak)^2 e^{-2kd} \tag{8}$$

In shallower water, Stokes drift contributes to sediment transport onto beaches and along beaches when the waves approach the coast at an angle. *See also:* NEARSHORE PROCESSES.

Shoaling and breaking

Waves approaching the shore from the open ocean are affected in several ways. As they become shallow-water waves, velocity C decreases with decreasing water depth h [Eq. (**4**)]. Consequently, waves approaching the shore at an angle are refracted so their crests are brought nearly parallel to the shoreline. Wave period T does not change as h decreases, so wavelength $L = CT$ must decrease according to Eq. (**4**). Shallowing also causes a growth in wave height as in relation (**9**),

$$H \propto h^{-r} \tag{9}$$

where the value of r depends on the character of the waves. For example, in the absence of dissipation, $r = 1/4$ according to linear theory for shallow-water waves [Eqs. (4)–(7)], while $r = 1$ for solitary waves of moderate amplitude.

As wave height grows, nonlinear terms in the equations of motion become significant. At first, this results in a vertical asymmetry in the wave shape, with crests becoming more peaked and troughs more rounded. Then, as the wave moves into shallow water, a strong horizontal asymmetry develops in which the forward face of the wave becomes progressively steeper than the backward face. This process typically continues until the wave

breaks. The resulting breakers can have a variety of forms (collapsing, plunging, surging and so forth), depending on the height of the entering waves and the slope of the seabed. A rough criterion for breaking is shown in relation (**10**),

$$H > \frac{C^2}{g} \tag{10}$$

or, using Eq. (**4**) for shallow water, $H > h$.

In the open ocean also, waves often break (**Fig. 1**). In this case, the criterion given by relation (10) becomes $H > 1/k$ [using Eq. (**4**) for deep water]. Whitecaps from deep-water breaking waves begin to appear at wind speeds of about 0.45 m/s (10 mi/h), while at wind speeds above 27 m/s (60 mi/h) all the high waves are breaking.

Wave measurement

There are three classes of instruments for measuring ocean surface waves: those which are at the air-sea interface, those which are below it and those which are above it. Because so many techniques can be used, only a few typical examples of each class will be described.

At the surface. Ocean surface waves can be measured from a dock with a wave staff held vertically in the water. The varying position $\eta(t)$ of the sea surface on the staff is sensed in a variety of ways, for example, by seawater-shorting of the submerged part of a resistance wire wound along the staff.

An accelerometer-instrumented buoy on a slack mooring line can provide a record of $\eta(t)$ by double integration of the vertical acceleration. If, in addition, the buoy contains tilt sensors, it is capable of providing the directional spectrum $S(f,\theta)$ of the waves. Typically, data from a wave-measuring buoy are telemetered to a receiving station on shore. *See also:* INSTRUMENTED BUOYS.

Below the surface. The most common instrument of this class is the subsurface pressure sensor. The pressure measurement must be made at a level deep enough that it is always submerged; this has the advantage of reducing vulnerability to damage by ships and breaking waves. The main disadvantage of the method is the need to compensate for the frequency-dependent depth attenuation of the measured wave-induced pressure signal $p(t)$ in converting it to surface elevation $\eta(t)$.

A narrow-beam inverted echo sounder can also be used to make subsurface wave measurements. It is placed on the seafloor and directed upward, so the acoustic echo time from the surface is a measure of sea-surface elevation. But variations in temperature and salinity affect the speed of sound in water and hence affect instrument calibration. Also, bubbles in the water can cause spurious acoustic reflections. *See also:* ECHO SOUNDER.

Above the surface. Ground-based high-frequency (3–30 MHz) radar systems can provide information on wave height and direction from the backscattered signal. Ranges of 50–500 km (30–300 mi) are feasible, but with over-the-horizon sky-wave systems relying on ionospheric reflection, ranges beyond 3200 km (2000 mi) have been achieved.

From aircraft, stereo-photographs can be taken of the sea surface and analyzed photogrammetrically, but this is a laborious process. Also, laser or narrow-beam radar ranging may be used to measure profiles of the sea surface. *See also:* LIDAR; PHOTOGRAMMETRY; RADAR.

Two radar techniques are presently used for wave measurements from satellites. The radar altimeter (13.5 GHz) observes the reflection of pulses directed vertically. The deformation of the reflected pulse carries information about significant wave height $H_{1/3}$ in the irradiated patch of ocean (several kilometers in diameter). The synthetic-aperture radar (SAR; >1 GHz) obliquely irradiates a patch of ocean surface (about 100 km or 60 mi in size) and uses pulse timing and phase information in the backscattered signal to obtain spatial resolution approaching 10 m (30 ft). The backscattering is from sea-surface roughness on scales comparable to the radar wavelength (several centimeters or inches), but longer-wavelength components appear in the SAR image by hydrodynamic interaction, electromagnetic modulation and effects of wave motion. As a result, it appears that directional wave spectra $S(f,\theta)$ may be obtained from SAR images, but the procedure is not yet fully understood. *See also:* METEOROLOGICAL SATELLITES.

Mark Wimbush

Sea state

Sea state is the description of the ocean surface or state of the sea surface with regard to wave action. Wind waves in the sea are of two types: those still growing under the force of the wind are called sea: those no longer under the influence of the wind that produced them are called swell. Differences between the two types are important in forecasting ocean wave conditions.

Sea. Those waves which are still growing under the force of the wind have irregular, chaotic and unpredictable forms. The unconnected wave crests are only two to three times as long as the distance between crests and commonly appear to be traveling in different directions, varying as much as 20° from the dominant direction. As the waves grow, they form regular series of connected troughs and crests with wave lengths commonly ranging from 12 to 35 times the wave heights. Wave heights only rarely exceed 55 ft (17 m). The appearance of the sea surface is termed state of the sea (**Table 1**).

The height of a sea is dependent on the strength of the wind, the duration of time the wind has blown and the fetch (distance of sea surface over which the wind has blown).

Swell. As sea waves move out of the generating area into a region of weaker winds, a calm, or opposing winds, their height decreases as they advance, their crests become rounded and their surface is smoothed. These waves

TABLE 1. Sea height code*

Code	Height, ft†	Description of sea surface
0	0	Calm, with mirror-smooth surface
1	0–1	Smooth, with small wavelets or ripples with appearance of scales but without crests
2	1–3	Slight, with short pronounced waves or small rollers; crests have glassy appearance
3	3–5	Moderate, with waves or large rollers; scattered whitecaps on wave crests
4	5–8	Rough, with waves with frequent whitecaps; chance of some spray
5	8–12	Very rough, with waves tending to heap up; continuous whitecapping; foam from whitecaps occasionally blown along by wind
6	12–20	High, with waves showing visible increase in height, with extensive whitecaps from which foam is blown in dense streaks
7	20–40	Very high, with waves heaping up with long frothy crests that are breaking continuously; amount of foam being blown from the crests causes sea surface to take on white appearance and may affect visibility
8	40+	Mountainous, with waves so high that ships close by are lost from view in the wave troughs for a time; wind carries off crests of all waves, and sea is entirely covered with dense streaks of foam; air so filled with foam and spray as to affect visibility seriously
9		Confused, with waves crossing each other from many and unpredictable directions, developing complicated interference pattern that is difficult to describe; applicable to conditions 5-8

* After *Instruction Manual for Oceanographic Observations*, H. O. Publ. 607, 2d ed., U.S. Navy Hydrographic Office, 1955.
† 1 ft = 0.3 m.

are more regular and more predictable than sea waves and, in a series, tend to show the same form or the same trend in characteristics. Wave lengths generally range from 35 to 200 times wave heights.

The presence of swell indicates that recently there may have been a strong wind, or even a severe storm, hundreds or thousands of miles away. Along the coast of southern California long-period waves are believed to have traveled distances greater than 5000 mi (8000 km) from generating areas in the South Pacific Ocean. Swell can usually be felt by the roll of a ship, and, under certain conditions, extremely long and high swells in a glassy sea may cause a ship to take solid water over its bow regularly.

When swell is obscured by sea waves, or when the components are so poorly defined that it is impossible to separate them, it is reported as confused (**Table 2**).

In-between state. Often both sea waves and swell waves, or two or more systems of swell, are present in the same area. When waves of one system are superimposed upon those of another, crests may coincide with crests and accentuate wave height, or troughs may coincide with crests and cancel each other to produce flat zones. This phenomenon is known as wave interference and the wave forms produced are extremely irregular. When wave systems cross each other at a considerable angle, the apparently unrelated peaks and hollows are known as a cross sea.

Breaking waves. The action of strong winds (greater than 12 knots or 6.2 m/s) sometimes causes waves in deeper water to steepen too rapidly. As the height-length ratio becomes too large, the water at the crest moves faster than the crest itself and topples forward to form whitecaps.

TABLE 2. Swell-condition code*

Code	Description	Height, ft†	Length, ft†
0	No swell	0	0
	Low swell	1–6	
1	Short or average		0–600
2	Long		600+
	Moderate swell	6–12	
3	Short		0–300
4	Average		300–600
5	Long		600+
	High swell	12+	
6	Short		0–300
7	Average		300–600
8	Long		600+
9	Confused		

*After *Instruction Manual for Oceanographic Observations*, H. O. Publ. 607, 2d ed., U.S. Navy Hydrographic Office, 1955.
†1 ft = 0.3 m.

As waves travel over a gradually shoaling bottom, the motion of the water is restricted and the wave train is telescoped together. The wave length decreases and the height first decreases slightly until the water depth is about one-sixth the deep-water wave length and then rapidly increases until the crest curves over and plunges to the water surface below. Swell coming into a beach usually increases in height before breaking, but wind waves are often so steep that there is little if any increase in height before breaking. For this reason, swell that is obscured by wind waves in deeper water often defines the period of the breakers.

The zone of breakers, or surf, includes the region of white water between the outermost breaker and the waterline on the beach. If the sea is rough, it may be impossible to differentiate between the surf inshore and the whitecaps in deep water just beyond.

Neil A. Benfer

Capillary waves

Capillary waves, or ripples, occur at the interface between two fluids when the principal restoring force is surface tension. Ripples generated by wind on the ocean and lakes are important for the initiation of turbulence in both media, transfer of gases between air to water and scattering of electromagnetic and sound waves.

Capillary waves have short wave lengths. Both the phase and group velocities increase as wave lengths become shorter and group velocity is greater than phase velocity. Dissipation of the waves by molecular viscosity is very rapid. The characteristic shape of the water surface, sinusoidal for small amplitudes, becomes distorted with more sharply curved troughs than crests as wave amplitude increases. In all these respects, ripples contrast sharply with gravity waves.

Mathematical formulas have been derived that relate the phase velocity, group velocity and frequency to the wave length of low-amplitude waves on still water in the absence of wind. In such waves, gravity and surface tension play equal roles in the restoring force; longer waves are considered to be gravity waves and shorter are capillary waves. *See also:* SURFACE TENSION.

In nature, ripples are observed to grow rapidly when the wind blows and to die away rapidly when the wind stops. When the water surface is uncontaminated, ripples die away to 37% of their original amplitude in a period of time that is related to the wave length and the kinematic viscosity of water. For example, for a wave with a wave length of 0.67 in. (17 mm) and a kinematic velocity of 10 m/s, the period of time is 3.8 s. When the water surface is contaminated, as by an oil film or other surface-active agent, ripples are damped still more rapidly, because the contaminated surface tends to act as an inextensible film against which the water motions due to the ripples must rub. In such a case for the example given, the period of time becomes 0.86 s.

Under low-wind conditions, this increased damping almost completely inhibits ripple growth: the surface appears glassy smooth and is called a slick. It has been observed that even short gravity waves grow at an inappreciable rate under such conditions; the interpretation here is that the fine scale of roughness presented to the wind by a rippled surface is involved in the formation and growth of gravity waves. However, ripples have been observed to form on clean water in the absence of wind by nonlinear processes occurring at the sharp crests of short, steep gravity waves; consequently the formation of both gravity waves and ripples is an interconnected process. Ripple trains formed under low-wind conditions derive their energy at the expense of gravity waves and are called parasitic capillaries. In this case, the capillary wave train is propagating in a moving stream of water, the orbital current of the underlying gravity wave. The longer wave parts of the capillary train would have a lower phase velocity in still water than either the shorter wave-length parts of the train or of the gravity wave, but they maintain their position relative to the crest of the gravity wave because they ride in a favorable part of the orbital current of the gravity wave. Short capillaries, having high phase velocity, do not need this aid to keep in phase and their higher group velocity enables them to proceed to a leading position where the gravity-wave orbital velocity vanishes or even opposes their motion.

Internal waves

Internal waves are wave motions of stably stratified fluids in which the maximum vertical motion takes place below the surface of the fluid. The restoring force is mainly due to gravity; when light fluid from upper layers is depressed into the heavy lower layers, buoyancy forces tend to return the layers to their equilibrium positions. Internal waves have been found in the atmosphere as lee waves (waves in the wind stream downwind from a mountain) and as waves propagated along an inversion layer (a layer of very stable air). They are also associated with wind shears at the lower boundary of the jet stream. In the oceans, internal oscillations have been observed wherever suitable measurements have been made. The observed oscillations can be analyzed into a spectrum with periods ranging from a few minutes to days. At a number of locations in the oceans, internal tides, or internal waves having the same periodicity as oceanic tides, are prominent.

Internal waves are important to the economy of the sea because they provide one of the few processes that can redistribute kinetic energy from near the surface to abyssal depths. When they break, they can cause turbulent mixing despite the normally stable density gradient in the ocean. Internal waves are known to cause time-varying refraction of acoustic waves because the sound velocity profile in the ocean is distorted by the vertical motions of internal waves. The result is that quasi-horizontal propagation of sound shows phase incoherence and large changes of intensity with time at ranges where the refraction has led to divergence or convergence of rays.

The vertical distribution of motions and phase velocity of internal waves depends on the vertical gradient of density in the fluid and the frequency of the generating forces. There is a simple density distribution that is illustrative: The fluid consists of two homogeneous layers, a lighter one on top of a heavier one, such as kerosene over water. The internal waves in this system are sometimes called boundary waves, because the maximum vertical motion occurs at the discontinuity of density at the boundary between the two fluids. Internal waves move at a slow speed, of the order of a few knots in the deep oceans. The effect of the rotation of the Earth is to increase the phase velocity of waves having periods long enough to approach one pendulum day.

When there is a continuous distribution of density in a fluid, as in the ocean or atmosphere, internal waves are possible only for frequencies that are lower than the maximum value given by a mathematical relationship known as the Väisälä-Brunt frequency, which is related to the downward rate of increase of the density and the velocity of sound in the fluid. In the ocean this maximum frequency occurs in the thermocline, where it commonly amounts to about one-fifth cycle per minute. At any frequency lower than this limit, there is an infinity of possible modes of internal waves. In the first mode, the vertical motion has a single maximum somewhere in the body of the fluid; in the second mode, there are two such maxima (180° out of phase), with a node between; and so on. The actual motion usually consists of a superposition of modes. In the ocean, the Väisälä-Brunt frequency varies from a maximum, commonly near 0.2 cycle per minute in the steepest part of the thermocline to negligible values at the bottom of the deep seas and in mixed layers such as those at the sea surface. Amplitudes of waves are oscillatory with depth and appreciable only where their frequency is less than the local value of the Väisälä-Brunt frequency. For this reason, internal waves of high frequency are limited in depth to the thermocline. Long-period (low-frequency) internal waves are affected by Earth rotation and the limiting periodicity is the local value of the pendulum day (one-half sidereal day divided by the sine of the latitude). These longest-period oscillations are inertial motions, in which the orbital path no longer has any vertical component but forms a circle in the horizontal.

In a continuously stratified fluid such as the ocean below the thermocline, internal waves propagate diagonally upward and downward, thus distributing energy throughout the ocean depths. Their ray paths are easily distorted because the group velocity of internal waves is comparable to the differences in velocity in the vertical of ocean currents and inertial motions. As a consequence, bundles of internal wave rays suffer severe refraction, sometimes to the extent that they form caustics known as critical layers, where the wave energy is absorbed by breaking. Where the sea bed is level, internal waves can be reflected by surface and seafloor to form normal modes within the body of the ocean. On a sloping seafloor, reflection at the bottom causes a change of wave length.

Internal waves in the atmosphere have been detected by a variety of instruments: microbarographs and wind recorders at ground level and long-term recordings of the scattering of radar or sonar beams by sharp density gradients in the high atmosphere. In the ocean, internal waves have been found by recording fluctuating currents in middepths by moored current meters, by acoustic backscatter Doppler methods and by studies of the fluctuations of the depths of isotherms as recorded by instruments repeatedly lowered from shipboard or by autonomous instruments floating deep in the water.

Internal waves are thought to be generated in the sea by variations of the wind pressure and stress at the sea surface, by the interaction of surface waves with each other and by the interaction of tidal motions with the rough seafloor. Their importance is that they can transmit energy and momentum throughout the ocean, not only laterally but also vertically. They can, therefore, transmit energy from the surface to all depths. In this way the otherwise sluggishly moving water at great depths can be agitated.

Charles S. Cox

Long-period waves

Long-period waves are those that exist when the period (T) is longer than one-half of a pendulum day, that is, 12 h/sin θ, where θ is the latitude of the location of interest. The main types of low-frequency ocean waves are Rossby waves, topographic Rossby waves, coastal Kelvin waves and equatorial Kelvin waves. *See also:* CORIOLIS ACCELERATION.

Rossby waves. Rossby waves, named after meteorologist C. Rossby, are fundamental to the large-scale dynamics of both atmosphere and ocean. They can exist at periods from a few days to several years and help to describe, for example, seasonal and climatic fluctuations in the oceans. Since they exist at long periods, the Earth rotates several times during a wave period and the rotation of the Earth therefore plays a central role in Rossby-wave dynamics. In order to understand Rossby waves, it is necessary to consider rotation, angular velocity and vorticity.

The angular velocity of the Earth is defined as a vector of magnitude Ω and direction northward along the axis of rotation. An angular velocity can be similarly defined for all solid rotating bodies: the magnitude of the angular velocity is 2π radians divided by the period of rotation and the direction is along the axis of rotation in the direction analogous to that for the rotating Earth. The vorticity of a particle of water in solid-body rotation is defined to be twice the angular velocity. In general, the velocity shear for a water particle will not correspond to that for solid-body rotation and the effective rotation and vorticity will consequently be modified. However, this does not affect the following explanation for the Rossby-wave mechanism.

Because the gravitational force perpendicular to the Earth's surface is so dominant, water particles tend to remain on the same horizontal level; that is, long-period ocean flows tend to be parallel to the Earth's surface. For this reason, rotational effects in the horizontal plane are of prime significance; consequently the vertical component of vorticity is of greatest importance to ocean dynamics. This vertical component of vorticity has two contributions.

One, which exists even when the water is at rest relative to the Earth, is the local vertical component of the Earth's rotation at latitude θ. The remaining contribution is due to rotational effects in the horizontal currents and its value is positive when rotation is counterclockwise as viewed locally above the Earth's surface.

Topographic Rossby waves. Another type of long-period ocean wave is the topographic Rossby wave, whose mechanism depends on variations of bottom topography and hence on the water depth. An example is the continental shelf wave in which water depth varies strongly perpendicular to the coastline across the continental shelf and slope. Continental shelf waves propagate with the coast on their right in the Northern Hemisphere and with the coast on their left in the Southern Hemisphere.

Like Rossby waves, continental-shelf waves also exist when the water is stratified. Then they are known as coastally trapped waves. These propagate along the coast in the same direction as shelf waves. Continental-shelf and coastally trapped waves play an important role in the dynamics of sea level and current fluctuations on the continental shelf. Much of the wind-forced ocean energy on the continental shelf is associated with these waves.

Coastal Kelvin waves. Coastal Kelvin waves, first discussed by Lord Kelvin, are low-frequency waves quite distinct from Rossby waves. For Kelvin waves, gravity is the essential restoring force; coastal Kelvin waves are long gravity waves trapped to a coastal wall by the Coriolis force. The amplitude of the wave decays exponentially from the coast. The water velocity perpendicular to the coast is zero everywhere. Like continental shelf waves, coastal Kelvin waves propagate with the coast on their right in the Northern Hemisphere and on their left in the Southern Hemisphere. The flow underneath a wave crest is in the direction of wave propagation, so the only way that the Coriolis force can balance the pressure-gradient force tending to flatten the sea surface is if the wave propagates with the coast on its right. Similar arguments show that propagation is with the coast on the left in the Southern Hemisphere.

Since a Kelvin wave is a long-period gravity wave, it can exist as either an external or internal gravity wave. In the external case, the wave behaves as though the water is of constant density and the phase speed (c) is given by the expression $\sqrt{gh}$, where g is the acceleration due to gravity and h is water depth. In the internal case, the phase speed also depends on the rate of density variation with depth. In general, internal Kelvin waves travel much more slowly than surface Kelvin waves.

Winds and tidal forces effectively generate oceanic coastal Kelvin waves. Strictly speaking, coastal Kelvin waves exist only when water of constant depth is bounded by a vertical wall and such topography in reality never occurs. However, for cases in which the decay scale is much greater than the distance across the shelf and slope to the constant-depth deep sea, the wave is dynamically coastal Kelvin. Such a condition is most often satisfied in the external Kelvin case.

Equatorial Kelvin waves. Equatorial Kelvin waves are long gravity waves with phase speed trapped to the Equator by the Coriolis force. The north-south amplitude variation is symmetric about the Equator and bell-shaped. The wave's north-south velocity is zero and the wave must propagate eastward. Equatorial Kelvin waves exist at very

low frequencies and the internal ones appear to play a fundamental role in the dynamics of climate fluctuations like those associated with El Niño. *See also:* EL NIÑO; OCEAN CIRCULATION.

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Additional Readings

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