

OCEAN WAVES

Waves-Introduction

[Waves](#) are the most conspicuous feature of the ocean. [Waves](#) sheer size and vigour have always impresses wave watchers. The scientific study of the [waves](#) began in the early 19th century. Franz Gerstner was the first who proposed to explain the phenomenon of [waves](#). According to Franz Gerstner water particles in a wave move in circular orbits. In 20th century oceanographers such as Harold U. Sverdrup and Walter munk of Scripps Institution of oceanography undertook a detailed study to predict wave and surf movements for naval operations.

The knowledge of wave is necessary for fishing off shore petroleum explorations, marine mining, marine engineering, and development of coastal areas.

Learning objectives

The present unit on ocean [waves](#) helps us to understand

1. What is wave
2. Types of wave
3. Theories of [wave formation](#)
4. Generation of different types of [waves](#)
5. Breaking and spilling [waves](#)
6. [Tsunamis](#) and its effects
7. [Internal waves](#)

Forces acting on seawater

According to George Karlenskit [waves](#) occur due to forces acting on the water surface. The energy that brings the wave is either dissipated at sea or strikes the beach. The [waves](#) result from distant geological effects and may travel thousands of miles before striking land. For instance the majority of the [waves](#) seen on ocean beaches result from distant winds and may range from small ripples to huge [tsunamis](#).

Waves

Definition

An ocean wave is an alternating rise and fall of a portion of the water surface or of a density boundary within the water column. [Waves](#) can be produced by the wind, movement of objects into or within the water and vibrations(earthquake) of the basin containing the water body. Production of a wave represents the transfer of energy into the water from the wave inducing source. Two basic types of [waves](#) are **progressive [waves](#)** and **standing [waves](#)** or **seiches**. A **progressive wave**: moves forward across the water from the area in which it was formed. **Seiches** remain where they are produced and appear as a rocking back-and-forth of the water surface or boundary layer.

Waves-terms and classification

a. **Catastrophic [waves](#)**: Sudden violent and temporary [waves](#) caused by earthquakes, volcanic activity etc are called catastrophic [waves](#). Catastrophic [waves](#) are produced with the sporadic occurrence of volcanoes, earthquakes or landslides in the ocean. These [waves](#) are generated less frequently, but they have more damaging effect. Three common types of catastrophic [waves](#) are [tsunamis](#), landslide surges and storm surges.

b. **Tsunamis**: These are seismic [waves](#) formed when a severe shock such as earthquake affects the ocean. They have wave length of 200Km and can travel as fast as 700 km/h.

c. Landslide surges: Rise or piling up of water against shore produced by land slide.

d. Storm surges: Storm surge is associated with weather and is very dangerous. The air pressure over a section of the ocean of the ocean affects the sea level. Sea level under a strong high pressure system is pushed downward to a level several centimeter below normal sea level. Conversely under an extreme low pressure, such as hurricane or tropical storm, a mound of water develops and is pushed along by the storm front. As the storm system approaches land the mound of sea water becomes a mass of wind driven, elevated water, usually associated with large storm [waves](#). Storm surges are more dangerous when they coincide with high [tides](#) they are responsible for the majority of flooding and destruction associated with hurricanes, 90% of people killed by hurricanes are killed by storm surges.

e. Shallow water [waves](#) : [Waves](#) in the water shallower than $L/2$.

f. Shearwave : [waves](#) that causes particles in a medium to vibrate back and forth at right angles to the direction of [waves](#) propagation.

g. Surface wave : [Waves](#) on the ocean surface formed at the boundary between two fluids of very different density. Surface wave have the property of reflection refraction and interference.

Wind produced [waves](#)

Three types of wind generated [waves](#) are sea, swell and surf. Most [waves](#) are formed by wind usually by storm systems. Unlike the storm systems over the land, ocean storm systems can be quite large, some exceeding 805Km in diameter. These systems break up as they approach land but over the ocean there is little to affect them. The wind transfers its energy to the water through wave building directly under the storm system is an area of mixed wave types termed as sea. Factors that affect the amount of energy transferred to the wave depend on wind speed, the duration of time that the wind blows in one direction and the "Fetch" the distance over which the wind blows in one direction. Fully developed sea is defined as the maximum size to which [waves](#) can grow under given conditions of wind speed, duration and fetch. At this point the [waves](#) of a fully developed sea will gain as much energy from the wind as they lost to gravity as breaking white caps.

- **Sea:** The wind transfers its energy to the water through wave building directly under the storm system is an area of mixed wave types termed as sea.
- **Swells:** As [waves](#) move away from the events that ruffle them, they assume a uniform pattern and begin to move in trains of equivalent period and height. These trains are called swells. Swells can travel thousands of miles.
- **Surf:** As swells approach shore their pattern is modified by shallow water and consequently they are shortened and as the crest increases, they move more swiftly. This modified form of swell is called surf. Surf is different from sea and swell i.e. It does not have the circular motion of wave particles.

a. Standing waves

Deep-water [waves](#), shallow-water [waves](#), and [internal waves](#) are all progressive [waves](#); they have a speed and move in direction. [Standing waves](#) do not progress; they are reflected back on themselves and appears as an alternation between a trough and a crest at a fixed position. They occur in ocean basins, partly enclosed bays, and estuaries. A standing wave can be demonstrated by slowly lifting one end of a container partially filled with water and then rapidly but gently returning it to a level position. If this is done, the surface alternately rises at one end and falls at the other end of the container. The surface oscillates about a point at the center of the container is the node, and the alternations of low and high water at each end are antinodes. A standing wave is a progressive wave reflected back on itself; the reflection

cancels out the forward motions of the initial and reflected [waves](#).

[Standing waves](#) in bays or inlets with an open end behave somewhat differently than [standing waves](#) in closed basins. A node is usually located at the entrance to the open-ended bay, so only one-quarter of the wavelength is inside the bay. There is little or no rise and fall of the water surface at the entrance, but a large rise and fall occurs at the closed end of the bay. Multiple nodes may also be present in open-ended basins.

Wave Parameters

All [waves](#) possess various features that can be measured. These include wave length, Waveheight, amplitude, Waveperiod and Wavefrequency. From these variables, wave stability and speed may be calculated.

Wave crest is the highest point on a wave and **wave trough** is the lowest point on a wave. **Wave length (L or λ)** is the length of one complete wave form, as measured from wave crest to adjacent wave crest (trough to trough). **Wave height (H)** is the vertical distance from wave crest to wave trough. **Wave Amplitude (A)** is the vertical distance of the crest or trough from the flat, undisturbed water level. It is equal to one-half the wave height ($A = 1/2 H$).

Wave period (T) is the amount of time required for one wave length to pass a fixed point. **Wave frequency (F)** is the number of [waves](#) passing a fixed point in a unit time and it is equal to one divided by the period ($F = 1/T$). **Celerity (C)** is the velocity with which a wave form travels. Celerity can be measured directly or calculated by dividing the wave length by wave period ($C = \lambda / T$). **Wave stability** or steepness is the ratio of wave height to wave length ($\text{stability} = H/\lambda$). A wave becomes unstable and collapses or breaks when $H/\lambda > 1/7$.

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Difference between surface and long [waves](#)

Surface [waves](#) are wind driven [waves](#), they are also called capillary [waves](#). These are tiny [waves](#) that result from forces of wind friction and surface tension. In the wind generated [waves](#) the surface tension provides the restoring force necessary to cause them to propagate. These [waves](#) have shorter wave lengths of about 1.73cm.

Longer [waves](#) have longer wave lengths and these [waves](#) travel faster than surface [waves](#). These are found in water deeper than $1/2$ their wavelengths and are also called deep water [waves](#).

Wave theories

[Linear Theory of Ocean surface waves](#): [Waves](#) are undulations of the sea surface with a height of around a meter, where height is the vertical distance between the bottom of a trough and the top of a nearby crest. The wavelength, which we might take to be the distance between prominent crests, is around 50m-100m. Watching the [waves](#) for a few minutes, one can notice that wave-height and wave-length are not constant. The heights vary randomly in time and space, and the statistical properties of the [waves](#), such as the average mean height for a few hundred [waves](#), change from day to day. These prominent offshore [waves](#) are generated by wind. Sometimes the local wind generates the [waves](#), other times distant storms generate [waves](#) which ultimately reach the coast. If we watch closely for a long time, one can notice that sea level changes from hour to hour. Over a period of a day, sea level increases and decreases relative to a point on the shore by about a meter. The slow rise and fall of sea level is due to the [tides](#), another type of wave on the sea

surface. [Tides](#) have wavelengths of thousands of kilometers, and they are generated by the slow, very small changes in gravity due to the motion of the sun and the moon relative to [earth](#).

Linear Theory of Ocean Surface waves

The ocean surface wave can be described quantitatively as follows. Surface [waves](#) are inherently nonlinear: The solution of the equations of motion depends on the surface boundary conditions.

By assuming that the amplitude of [waves](#) on the water surface is infinitely small so the surface is almost exactly a plane. To simplify the mathematics, one can also assume that the flow is 2-dimensional with [waves](#) travelling in the x-direction. We also assume that the Coriolis force and viscosity can be neglected.

With these assumptions, the sea-surface elevation ζ of a wave traveling in the x direction is :

With

$$\zeta = a \sin(kx - \omega t)$$

Where ω is wave frequency in radians per second, f is the wave frequency in Hertz(Hz) k is wave number, T is wave period, L is wave-length, and where we assume, as stated above, that $k a = 0$ (0)

The [wave period](#) T is the time it takes two successive wave crests or troughs to pass a fixed point. The wave-length L is the distance between two successive wave crests or troughs at a fixed time.

Dispersion Relation

Wave frequency ω is related to wave number K by the dispersion relation

$$\omega^2 = g k \tanh(kd)$$

Where d is the water depth and g is the acceleration of gravity

Two approximations are especially useful.

1. Deep-water approximation is valid if the water depth d is much greater than the wave-length L . In this case, $d \gg L$, $kd \gg 1$, and $\tanh(kd) = 1$

2. Shallow-water approximation is valid if the water depth is much less than a wavelength. In this case,

$$d \ll L, \text{ and } \tanh(kd) = kd$$

For these two limits of water depth compared with wavelength the dispersion relation reduces to:

$$\omega^2 = g k \quad \text{Deep-water dispersion relation}$$

$$d > L/4$$

$$\omega^2 = g k^2 \quad \text{Shallow-water dispersion relation}$$

$$d < L/11$$

The stated limits for d/L give a dispersion relation accurate within 10%. Because many wave properties can be measured with accuracies of 5-10%, the approximations are useful for calculating wave properties. Later we will learn to calculate wave properties as the [waves](#) propagate from deep to shallow water.

Phase velocity

The phase velocity c is the speed at which a particular phase of the wave propagates, for example, the speed of propagation of the wave crest. In one [wave period](#) T the crest advances one wave-length L and the phase speed is $c = L/T = \omega/k$.

Thus,, the definition of phase speed is:

$$C = \frac{\omega}{K}$$

The direction of propagation is perpendicular to the wave crest and toward the positive x direction. The deep-and shallow-water approximations for the [dispersion relation](#) give

Group velocity

The concept of group velocity C_g is fundamental for understanding the propagation of linear and nonlinear [waves](#). First, it is the velocity at which a group of [waves](#) travels across the ocean. More importantly, it is also the propagation velocity of [wave energy](#) which gives a clear derivation of the concept and the fundamental equation .

The definition of group velocity in two dimension is

$$C_g = \frac{\partial \omega}{\partial k}$$

Using the approximations for the [dispersion relation](#):

$$C_g = \frac{g}{2\omega} = \frac{c}{2}$$

[Deep water](#) group velocity

$$C_g = \frac{g}{2\omega} = c$$

Shallow-water group velocity

For ocean-surface [waves](#), the direction of propagation is perpendicular to the wave crests in the positive x direction. In the more general case of other types of [waves](#), such as **Kelvin and Rossby waves** the group velocity is not necessarily in the direction perpendicular to wave crests.

Notice that a group of deep-water [waves](#) moves at half the phase speed of the [waves](#) making up the group. How can this happen? If we could watch closely a group of [waves](#) crossing the sea, we would see [waves](#) crests appear at the back of the wave train, move through the train, and disappear at the leading edge of the group.

Each wave crest moves at twice the speed of the group. Do real ocean [waves](#) move in groups governed by the [dispersion relation](#)? Yes. Walter Munk and colleagues (1963) in a series of experiments in the 1960s showed that ocean [waves](#) propagating over great distances are dispersive, and that the dispersion could be used to track storms. They recorded [waves](#) for many days using an array of three pressure gauges just offshore of San Clemente Island, 60 miles due west of San Diego, California. Wave spectra were calculated for each day's data. From the spectra, the amplitudes and frequencies of the low-frequency [waves](#) and the propagation direction of the [waves](#) were calculated. Finally, they plotted contours of [wave energy](#) on a frequency-time diagram.

Contours of [wave energy](#) on a frequency-time plot calculated from spectra of [waves](#) measured by pressure gauges offshore of southern California. The ridges of high [wave energy](#) show the arrival of dispersed wave trains from distant storms. The slope of the ridge is inversely proportional to distance to the storm. Δ is distance in degrees, θ is direction of arrival of [waves](#) at California. From Munk et al. (1963).

To understand the figure, consider a distant storm that produces [waves](#) of many frequencies. The lowest-frequency [waves](#) (smallest W) travel the fastest, and they arrive before other, higher –frequency [waves](#). The further away the storm the longer the delay between arrivals

of [waves](#) of different frequencies. The ridges of high [wave energy](#) seen in the figure degrees Δ along a great circle; and the phase information from the array gives the angle to the storm. The two angles give the storm's location relative to San Clemente. Thus [waves](#) arriving from 15 to 18 September produce a ridge indicating the storm was 115° away at an angle of 205° which is south of New Zealand near Antarctica.

The locations of the storms producing the [waves](#) recorded from June through October 1959 were compared with the location of storms plotted on weather maps and in most cases the two agreed well.

Wave Energy

Wave energy E in Joules per square meter is related to the variance of sea-surface displacement by:

$$E = \rho_w \omega^2 \langle \eta^2 \rangle$$

Where ρ_w is water density, g is gravity, and the brackets denote a time or space average.

Significant Wave-Height

What do we mean by wave-height? If we look at a wind-driven sea, one can [waves](#) of various heights. Some are much larger than most, others are much smaller. A [practical](#) definition that is often used is the height of the highest of the highest 1/3 of the [waves](#), $H_{1/3}$. The height is computed as follows: measure wave-height for a few minutes, pick out say 120 wave crests and record their heights. Pick the 40 largest [waves](#) and calculate the average height of the 40 values. This is $H_{1/3}$ for the wave record.

Wave height estimated by observers corresponds to the average of the highest 20 to 40 per cent of [waves](#).... Originally, the term significant wave-height was attached to the average of these observations, the highest 30 per cent of the [waves](#), but has evolved to become the average of the highest one-third of the [waves](#), (designated H_s to $H_{1/3}$)

More recently, significant wave-height is calculated from measured wave displacement. If the sea contains a narrow range of wave frequencies, $H_{1/3}$ is related to the standard deviation of sea-surface displacement

$$H_{1/3} = 4 \langle \eta^2 \rangle^{1/2}$$

Where $\langle \eta^2 \rangle^{1/2}$ is the standard deviation of surface displacement. This relationship is much more useful, and it is now the accepted way to calculate wave-height from [wave measurements](#).

Non-linear waves

One can derive the properties of an ocean surface wave assuming [waves](#) were infinitely small $ka \rightarrow 0$. If the [waves](#) are small $ka \ll 1$ but not infinitely small, the wave properties can be expanded in a power series of ka (Stokes, 1847). He calculated the properties of a wave of finite amplitude and found:

The phase of the components for the Fourier series expansion of η in above equation

$$\zeta = a \cos(kx - \omega t) + \frac{1}{2} \epsilon^2 \cos(2kx - 2\omega t) + \dots$$

are such that non-linear [waves](#) have sharpened crests and flattened troughs. The maximum amplitude of the Stokes wave is $a_{\max} = 0.07L$ ($ka = 0.44$). Such steep [waves](#) in [deep water](#) are called Stokes [waves](#) (See also Lamb, 1945, §250).

Knowledge of non-linear [waves](#) came slowly until Hasselmann (1961, 1963a, 1963b, 1966), using the tools of high-energy particle physics, worked out to 6th order the interaction of three or more [waves](#) on the sea surface. He, Phillips (1960), and Longuet-Higgins and Phillips

(1962) showed that n free [waves](#) on the sea surface can interact to produce another free wave only if the frequencies and wave numbers of the interacting [waves](#) sum to zero:

$$\omega_1 \pm \omega_2 \pm \omega_3 \pm \dots \pm \omega_n = 0 \quad \text{-- 2}$$

$$k_1 \pm k_2 \pm k_3 \pm \dots \pm k_n = 0 \quad \text{-- 3}$$

$$\omega_i^2 = g k_i$$

Where we allow [waves](#) to travel in any direction, as above equation 2 & 3 and k_i is the vector wave number giving wave-length and direction are general requirements for any interacting [waves](#). The fewest number of [waves](#) that meet the conditions of are three [waves](#) which interact to produce a fourth. The interaction is weak; [waves](#) must interact for hundreds of wave-lengths and periods to produce a fourth wave with amplitude comparable to the interacting [waves](#). The Stokes wave does not meet the criteria of and the wave components are not free [waves](#); the higher harmonics are bound to the primary wave.

Wave Momentum

The concept of wave momentum has caused considerable confusion. In general, [waves](#) do not have momentum, a mass flux, but they do have a momentum flux. This is true for [waves](#) on the sea surface. Ursell (1950) showed that ocean swell on a rotating [Earth](#) has no mass transport. His proof seems to contradict the usual textbook discussions of steep, [non-linear waves](#) such as Stokes [waves](#). Water particles in a **Stokes wave** move along earths that are nearly circular, but the paths fail to close, and the particles move slowly in the direction of wave propagation. This is a mass transport, and the phenomena is called Stokes drift. But the forward transport near the surface is balanced by an equal transport in the opposite direction at depth, and there is no net mass flux.

Solitary waves

Solitary [waves](#) are another class of [non-linear waves](#), and they have very interesting properties. They propagate without change of shape, and two solutions can cross without interaction. The first solution was discovered by John Scott Russell (1808-1882), who followed a solitary wave generated by a boat in Edinburgh's Union Canal in 1834.

The properties of a solitary [waves](#) result from an exact balance between dispersion which tends to spread the solitary wave into a train of [waves](#), and non-linear effects which tend to shorten and steepen the wave. The type of solitary wave in shallow water seen by Russell, has the form:

$$\Sigma = a \operatorname{sech}^h \left(\frac{3a}{4d^3} (x-ct) \right)^{1/2}$$

Which propagates at a speed:

$$C = C_0 \left(1 + \frac{a}{d} \right)^{-2d}$$

You might think that all shallow-water [waves](#) are solutions because they are non-dispersive, and hence they ought to propagate without change in shape. Unfortunately, this is not true if the wave have finite amplitudes. The velocity of the wave depends on depth. If the wave consists of a single hump, then the water at the crest travels faster than water in the trough, and the wave steepens as it moves forward. Eventually, the wave becomes very steep and breaks. At this point it is called a bore. In some river mouths, the incoming [tide](#) is so high and the [estuary](#) so long and shallow that the tidal wave entering the estuary eventually steepens and breaks producing a **bore** that runs up the river.

Chapter 3: Ocean [Waves-3](#)

Beaufort scale

Beaufort Wind Scale and Sea State Chart

Sea State	Beaufort Number	Terminology	Wind Speed (miles per hour)	Average Wave Height (feet)	Description
0	0	Calm	0-1	0	Sea like a mirror
1	1	Light air	1-3	0.005	Ripples with appearance of scales; no crests
2	2	Light breeze	4-7	0.18	Small wavelets; crests have glossy appearance but do not break.
3	3	Gentle breeze	8-12	0.19-1	Large wavelets, crests begin to break; foam of glossy appearance; a few white horses.
	4	Moderate breeze	13-18	2-3	Small waves becoming longer form; many white horses formed; some spray
4	5	Fresh breeze	19-24	4-5	Moderate waves taking longer form; many white horses formed; some spray
	6	Strong breeze	25-31	6-10	Large waves begin to form; many white crests more spray
	7	Moderate gale	32-38	11-16	Sea heaps up; white foam from breaking waves begins to be blown in streaks
5	8	Fresh gale	39-46	17-28	Moderately high waves of greater length; edges of crests begin to break into spindrift; foam blown in well marked streaks

					spray affects visibility.
	9	Strong gale (storm)	47-54	29-40	High waves ; dense streaks of foam along wind direction; sea begins to roll; visibility affected
	10	Whole gale (heavy storm)	55-63	41-59	Very high waves with long overhanging crests; sea takes white appearance; rolling heavy; visibility affected.
	11	Violent storm	64-72	60-73	Exceptionally high waves ; small and medium sized ships lost from view; crests blown into froth.
	12	Hurricane	73-82	74-80	Air filled with foam and spray; sea completely white with driving spray; visibility very seriously affected.

Breaking waves

Breaking [waves](#) are otherwise called breakers. These are found in all beaches. Some [waves](#) are destroyed by opposite winds, others interact and some cancel each other, and most of the [waves](#) end up as breaker, when they encounter shallow water. Most [waves](#) move across the surface of the deep ocean unaffected by the bottom. As [waves](#) approach the coast, they are increasingly affected by the bottom, and they change from deep water to shallow water [waves](#). Wave length and speed decrease, while the wave period remains constant. At the same time the wave height decreases slightly and the finite energy is converted into potential energy and its height increases rapidly as water depth decrease to one tenth the wave length and the wave crusts found close together. [The crust](#) angle reach the critical value of 120° . The proximity of the causes once circular orbit to become elliptical further adding to the [waves](#) in stability. In deep water a molecules in [the crust](#) stop its orbit can count on support from the water immediately in front of it to complete its cycle. As the breaker shoals, there is not enough water immediately in front of [the crust](#) to come all the way up to meet it. [The crust](#) appears to get ahead of its support and it breaks. The wave break when the velocity of the water molecule in [the crust](#) is greater than that of the wave beneath such that [the crust](#) gets ahead of the wave.

[Waves](#) often break more than once before reaching shore. Breaker are classified in to four kinds

Types of breakers:

1. Spilling breakers

2. Plunging breakers
3. Collapsing breakers
4. Surging breakers

Spilling breakers: These are over steepened [waves](#) where the unstable top spills over the front of [waves](#) as they move towards a beach. In spilling breakers waveforms advance but wave heights are diminished. These [waves](#) break gradually over a considerable distance.

Plunging breakers: The wave crest typically curls over, forming a large airpacket, when a wave breaks, there is a large splash of water and foam which is usually thrown into the air. These plunging breakers tend to form from long gently sloping but irregular bottom.

Collapsing breaker: Break in the middle or near the bottom of the wave rather than at the top.

Surging breaker: In surging breaker the [waves](#) slides up and down the beach with little or no bubble production. These are common in areas with steeply sloping bottom.

Chapter 4: Ocean [Waves](#)-4

Tsunamis

Tsunamis are triggered by earthquake, underwater volcanoes, land slides, and melting of icebergs from glacier. Sudden movement of the [earth](#)'s crust may also produce seismic sea [waves](#), or tsunamis. These [waves](#) are often incorrectly called tidal [waves](#). Since a seismic sea wave has nothing to do with [tides](#). Most tsunamis are in the Pacific Ocean and are not so common in the Atlantic and Indian Oceans. The Hawaiian Islands experience a tsunami every two years.

If a large area of several hundred square kilometers, of the [earth](#)'s crust below the sea surface is suddenly displaced, it causes a sudden rise or fall in the sea surface level above it. In the case of a rise, gravity causes the suddenly elevated water to return to the equilibrium surface level; if a depression is produced, gravity forces cause the surrounding water to flow into it. Both cases result in the production of [waves](#) with extremely long wave lengths (100 to 200 km) and long periods as (10-20 minutes). Since the average depth of the oceans is about 4000m (or 4 km or 13,000 ft.) this depth is less than one-twentieth the wavelength of these [waves](#), and tsunamis are shallow-water [waves](#). These seismic [waves](#) gD and about 200 m/sec (400mbp). Since they are shallow-water \sqrt{gD} radiate from the point of the seismic disturbance at gD and move across the oceans at a speed determined by the ocean's depth ($C = \sqrt{gD}$), tsunamis may be refracted, diffracted, or reflected in midocean.

When a tsunami leaves its point of origin, it may have a height of 1 to 2 m, but this height is distributed over its many-kilometer wavelength. It is not easily seen or felt when superimposed on the other distortions of the sea's surface, and a vessel in the open ocean is in little or no danger if a tsunami passes. The danger occurs only if the vessel has the misfortune to be directly above the area the original seismic disturbance.

The energy of a tsunami is distributed from the ocean surface to the [ocean floor](#) and over the length of the wave. When the path of the wave is blocked by a coast or island, the wave behaves like any other shallow-water wave, and the energy is compressed into a smaller water volume as the depth rapidly decreases. This rapid and sudden increase in energy density causes the wave height to build rapidly, and the loss of energy is also rapid when the wave breaks. A tremendous surge of moving water races up over the land, destroying buildings, docks, and trees.

A tsunami often affects an area with a radius of more than 150km in exceptional circumstance and it can have a wave length as high as 1000km .when a tsunami hits the coast. It comes in [waves](#) after [waves](#) and in most cases the third to eighth [waves](#) are the most dangerous as they are the largest.

The leading edge of the tsunami wave group may be either a crest or a trough. If the initial crustal disturbance was an upward motion, a crest is formed first; if the crustal motion was downward, a trough is formed. If a trough arrives before the first crest, sea level drops rapidly, exposing sea plants and animals.

Tsunami prediction is an important and difficult task. The detection of under water seismic activity somewhere in the world does not mean that a tsunami has been generated. Because the [waves](#) are so long and low. They cannot be detected until they come into shore somewhere. Even though a tsunami is detected on some coast it is not possible to predict what other coastal area might be hit. Due to refraction by [sea floor](#) topography, these seismic [waves](#) will go some direction and not other. As they move at a speed of about 700km/hour across deep water it may not be possible to warn or evacuate coastal population even if one did know where the [waves](#) are headed.

Seiches

[Standing waves](#) that occur in natural basins are called **seiches**, and the oscillation of the surface is called **seiching**. In natural basins, the length dimension usually greatly exceeds the depth. Therefore, a standing wave of one node in such a basin behaves as a reflecting shallow-water wave, with the wave length determined by the length of the basin. In water with distinct layers having sharp density boundaries, [standing waves](#) may occur along the fluid boundaries as well as at the air-sea boundary. The oscillation of the internal [standing waves](#) is slower than the oscillation of the sea surface.

[Standing waves](#) may be triggered by tectonic movements that suddenly hit the basin, causing the water to oscillate at a period defined by the dimensions of the basin. If storm winds create a change in surface level to produce storm surges, the surface may oscillate as a standing wave in the act of returning to its normal level when the wind ceases. The pulsing of a weather disturbance over a lake may also cause periodic water-level change, reaching a meter or more in height. If the period of the disturbing force is a multiple of the natural period of oscillation of the basin, the height of the standing wave is greatly increased.

Internal waves

Below the ocean surface, [waves](#) form along the density boundaries between water layers of different density just as they do between air and water. These [waves](#) are called **internal waves**.

Internal [waves](#) form at the interfaces between layers of different water density. The disturbance needed to begin an internal wave need not be large, small generating forces plus the viscosity of the water results in large-amplitude, slow-moving [waves](#) with long wavelengths called internal [waves](#). Internal [waves](#) appear to move like shallow water [waves](#), with the thickness of the layers playing a role similar to the water depth in surface shallow-water [waves](#). Internal wave travel more slowly than surface [waves](#) of similar amplitude.

Internal [waves](#) may be caused when the surface layer flows over a more static lower layer in the same way the wind produces surface [waves](#). The possible triggers of internal [waves](#) include seismic disturbances, long-term wind changes, and tidal forces.

When the surface layer is relatively shallow, a wave moving along an internal boundary may be high enough to have its crest below the surface. This action creates moving bands of smooth water called slicks, each representing a crest position. Slicks are most likely to occur in coastal water, estuaries, or fjords, where the shallow surface layer is diluted by fresh water runoff. These natural slicks reveal the presence of underlying internal [waves](#).

Since internal [waves](#) occur below the surface and are usually not visible, instruments that

measure the periodic rise and fall of surfaces of constant temperature and salinity record the passing of internal [waves](#). Because they cannot be observed easily and are not common surface [waves](#). The speed of the [waves](#) depends on the differences in density between the two fluids. Surface wave can be up to 20m high, while internal [waves](#) can reach a height of 300m or more dependent on the thickness of the upperwater layer. Internal [waves](#) are of importance in the context of vertical mixing process, internal [waves](#) may transport plankton and fish larvae to other areas.