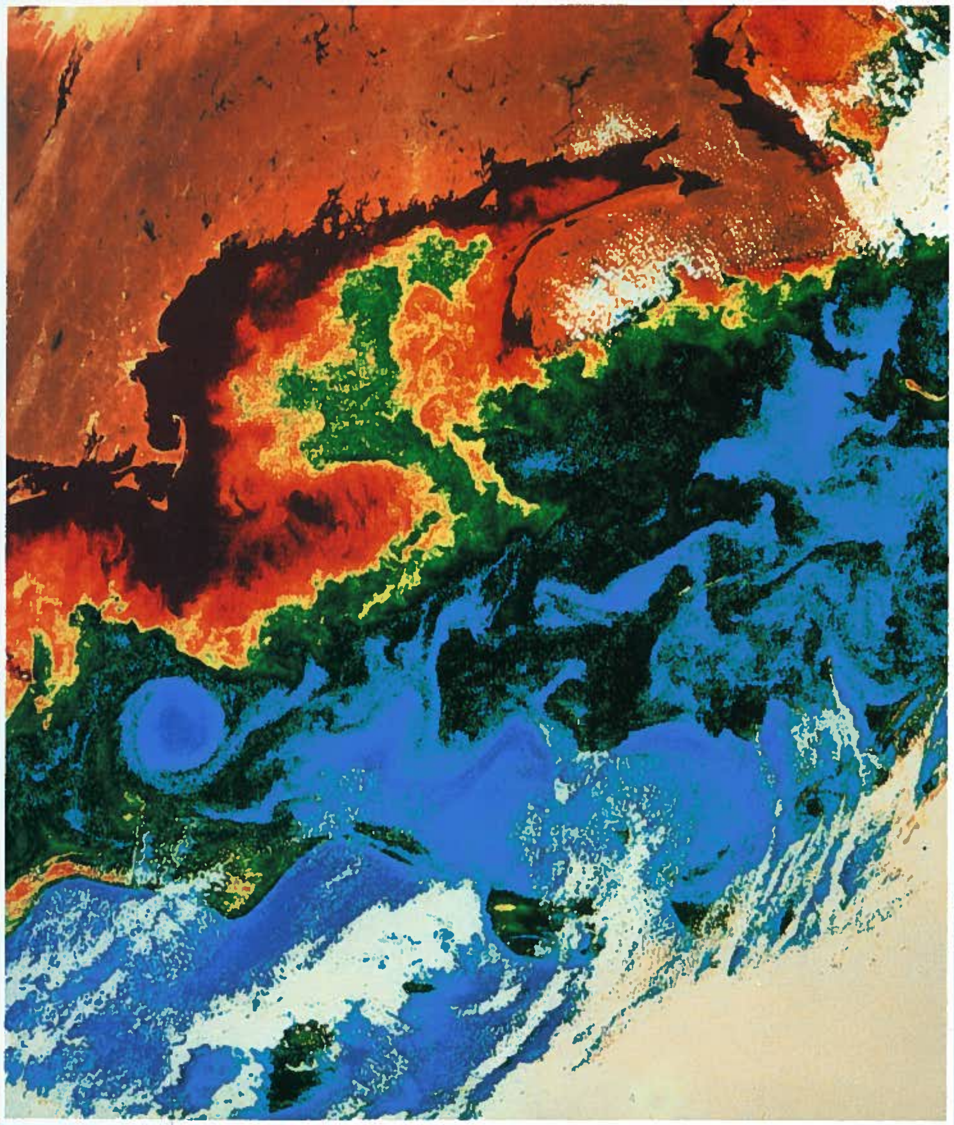
WAVES, TIDES AND SHALLOW-WATER PROCESSES

SECOND EDITION

PREPARED BY AN OPEN UNIVERSITY COURSE TEAM



The Open University

CHAPTER 1 WAVES

'... the chidden billow seems to pelt the clouds ...'

Othello, Act II, Scene I.

Sea waves have attracted attention and comment throughout recorded history. Aristotle (384–322 BC) observed the existence of a relationship between wind and waves, and the nature of this relationship has been a subject of study ever since. However, at the present day, understanding of the mechanism of wave formation and the way that waves travel across the oceans is by no means complete. This is partly because observations of wave characteristics at sea are difficult, and partly because mathematical models of wave behaviour are based upon the dynamics of idealized fluids, and ocean waters do not conform precisely with those ideals. Nevertheless, some facts about waves are well established, at least to a first approximation, and the purpose of this Chapter is to outline the qualitative aspects of water waves and to explore some of the simple relationships of wave dimensions and characteristics.

We start by examining the dimensions of an idealized water wave, and the terminology used for describing waves (Figure 1.1).

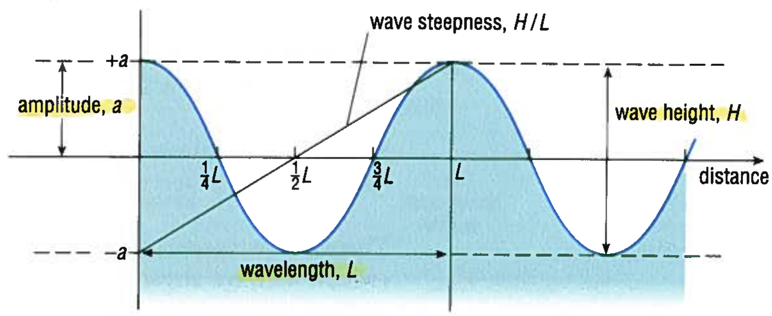


Figure 1.1 Vertical profile of two successive idealized ocean waves, showing their linear dimensions and sinusoidal shape.

Wave height (*H*) refers to the overall vertical change in height between the wave crest (or peak) and the wave trough. The wave height is twice the wave **amplitude** (*a*). **Wavelength** (*L*) is the distance between two successive peaks (or two successive troughs). **Steepness** is defined as wave height divided by wavelength (*H*/*L*) and, as can be seen in Figure 1.1, is not the same thing as the slope of the sea-surface between a wave crest and its adjacent trough. The time interval between two successive peaks (or two successive troughs) passing a fixed point is known as the **period** (*T*), and is generally measured in seconds. The number of peaks (or the number of troughs) which pass a fixed point per second is known as the **frequency** (*f*).

QUESTION 1.1 If a wave has a frequency of 0.2 s^{-1}, what is its period?

As the answer to Question 1.1 shows, period is the reciprocal of frequency. We will return to this concept in Section 1.2.

1.1 WHAT ARE WAVES?

Waves are a common occurrence in everyday life, and are manifested as, for example, sound, the motion of a plucked guitar string, ripples on a pond, or the billows on the ocean. It is not easy to define a wave. Before attempting to do so, let us consider some of the characteristics of wave motion:

1 A wave transfers a disturbance from one part of a material to another. (The disturbance caused by dropping a stone into a pond is transmitted across the pond by ripples.)

2 The disturbance is propagated through the material without any substantial overall motion of the material itself. (A floating cork merely bobs up and down on the ripples, but experiences very little overall movement in the direction of travel of the ripples.)

3 The disturbance is propagated without any significant distortion of the wave form. (A ripple shows very little change in shape as it travels across a pond.)

4 The disturbance appears to be propagated with constant speed.

If the material itself is not being transported by wave propagation, then what *is* being transported?

The answer, 'energy', provides a reasonable working definition of wave motion – a means whereby energy is transported across or through a material without any significant overall transport of the material itself.

So, if energy, and not material, is being transported, what is the nature of the movement observed when ripples cross a pond?

There are two aspects to be considered: first, the progress of the waves (which we have already noted), and secondly, the movement of the water particles themselves. Superficial observation of the effect of ripples on a floating cork suggests that the water particles move 'up and down', but closer observation will reveal that, provided the water is very much deeper than the ripple height, the cork is describing a nearly circular path in a vertical plane, parallel with the direction of wave movement. In a more general sense, the particles are displaced from an equilibrium position, and a wave motion is the propagation of regular oscillations about that equilibrium position. Thus, the particles experience a displacing force and a restoring force. The nature of these forces is often used in the descriptions of various types of waves.

1.1.1 TYPES OF WAVES

All waves can be regarded as **progressive waves**, in that energy is moving through, or across the surface of, the material. The so-called **standing wave**, of which the plucked guitar string is an example, can be considered as the sum of two progressive waves of equal dimensions, but travelling in opposite directions. We examine this in more detail in Section 1.6.4.

Waves which travel through the material are called body waves. Examples of body waves are sound waves and seismic P- and S-waves, but our main concern in this Volume is with *surface waves* (Figure 1.2). The most familiar surface waves are those which occur at the interface between atmosphere and ocean, caused by the wind blowing over the sea. Other external forces acting on the fluid can also generate waves. Examples range from raindrops falling into tidal pools, through diving gannets and ocean-going liners to earthquakes (see Section 1.6.3).

The tides are also waves (Figure 1.2), caused by the gravitational influence of the Sun and Moon and having periods corresponding to the causative forces. This aspect is considered in more detail in Chapter 2. Most other waves, however, result from a non-periodic disturbance of the water. The water particles are displaced from an equilibrium position, and to regain that position they require a restoring force, as mentioned above. The restoring force causes a particle to 'overshoot' on either side of the equilibrium position. Such alternate displacements and restorations establish a characteristic oscillatory 'wave motion', which in its simplest form has sinusoidal characteristics (Figures 1.1 and 1.6), and is sometimes referred to as simple harmonic motion. In the case of surface waves on water, there are two such restoring forces which maintain wave motion:

1 The gravitational force exerted by the Earth.

2 Surface tension, which is the tendency of water molecules to stick together and present the smallest possible surface to the air. So far as the effect on water waves is concerned, it is as if a weak elastic skin were stretched over the water surface.

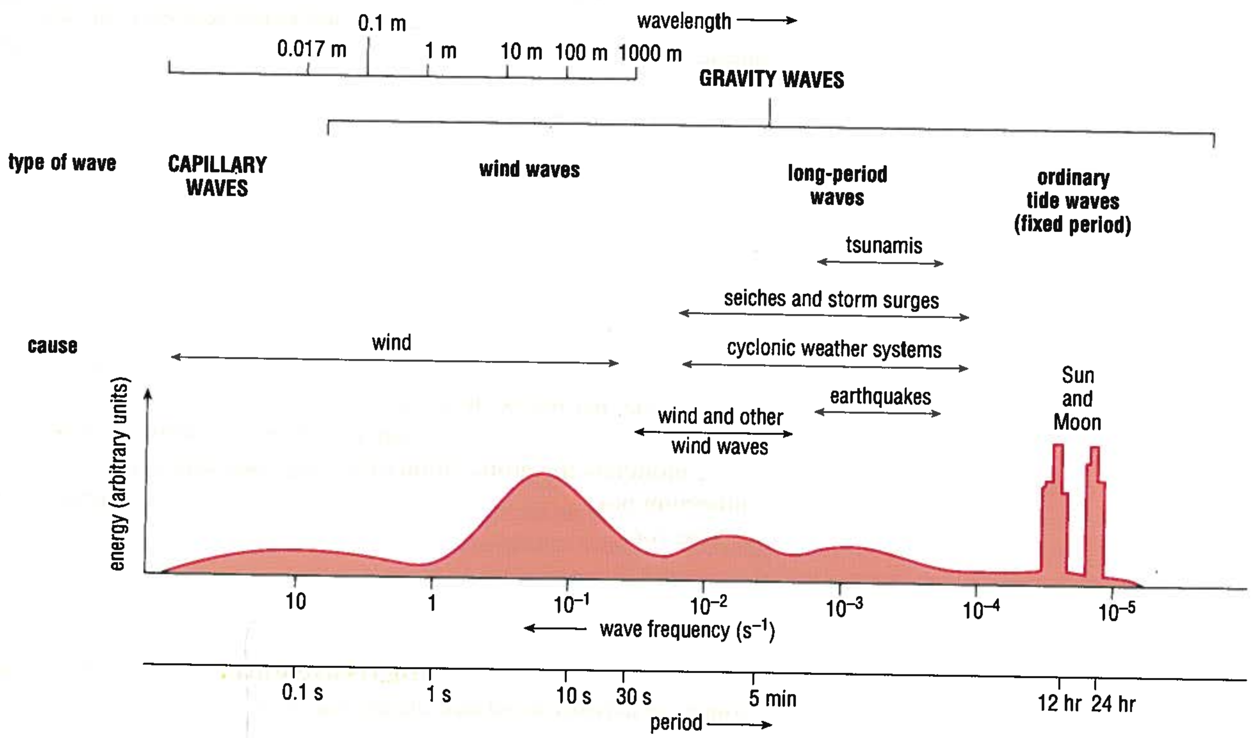


Figure 1.2 Types of surface waves, showing the relationships between wavelength, wave frequency and period, the nature of the forces that cause them, and the relative amounts of energy in each type of wave. Unfamiliar terms will be explained later. *Note:* Waves caused by 'other wind waves' are waves resulting from interactions between waves of higher frequency as they move away from storm areas – see Section 1.4.2.

Water waves are affected by both of these forces. In the case of waves with wavelengths less than about 1.7 cm, the principal restoring force is surface tension, and such waves are known as **capillary waves**. They are important in the context of remote sensing of the oceans (Section 1.7.1). However, the main interest of oceanographers lies with surface waves of wavelengths greater than 1.7 cm, and the principal restoring force for such waves is gravity; hence they are known as **gravity waves** (Figure 1.2).

Gravity waves can also be generated at an interface between two layers of ocean water of differing densities. Because the interface is a surface, such waves are, strictly speaking, surface waves, but oceanographers usually refer to them as **internal waves**. These occur most commonly where there is a rapid increase of density with depth, i.e. a steep density gradient, or **pycnocline**. Pycnoclines themselves result from steep gradients of temperature and/or salinity, the two properties which together govern the density of seawater. Because the difference in density between two water layers is much smaller than that between water and air, less energy is required to displace the interface from its equilibrium position, and oscillations are more easily set up at an internal interface than at the sea-surface. Internal waves travel considerably more slowly than most surface waves. They have greater amplitudes than all but the largest surface waves (up to a few tens of metres), as well as longer periods (minutes or hours rather than seconds, cf. Figure 1.2) and longer wavelengths (hundreds rather than tens of metres). Internal waves are of considerable importance in the context of vertical mixing processes in the oceans, especially when they break.

Not all waves in the oceans are displaced primarily in a vertical plane. For example, because atmosphere and oceans are on a rotating Earth, variation of **planetary vorticity** with latitude (i.e. variation in the angular velocity of the Earth's surface and hence in the effect of the Earth's rotation on horizontal motions) causes horizontal deflection of atmospheric and oceanic currents, and provides restoring forces which establish oscillations mainly in a horizontal plane, so that easterly or westerly currents tend to swing back and forth about an equilibrium latitude. These large-scale horizontal oscillations are known as **planetary** (or **Rossby**) **waves**, and may occur as surface or as internal waves. They are not gravity waves (i.e. the restoring force is not gravity) and so do not appear in Figure 1.2.

1.1.2 WIND-GENERATED WAVES IN THE OCEAN

In 1774, Benjamin Franklin said: 'Air in motion, which is wind, in passing over the smooth surface of the water, may rub, as it were, upon that surface, and raise it into wrinkles, which, if the wind continues, are the elements of future waves'.

In other words, if two fluid layers having differing speeds are in contact, there is frictional stress between them and there is a transfer of momentum and energy. The frictional stress exerted by a moving fluid is proportional to the *square* of the speed of the fluid, so the **wind stress** exerted upon a water surface is proportional to the square of the wind speed. At the sea-surface, most of the transferred energy results in waves, although a small proportion is manifest as wind-driven currents. In 1925, Harold Jeffreys suggested that waves obtain energy from the wind by virtue of pressure differences caused by the sheltering effect provided by wave crests (Figure 1.3).

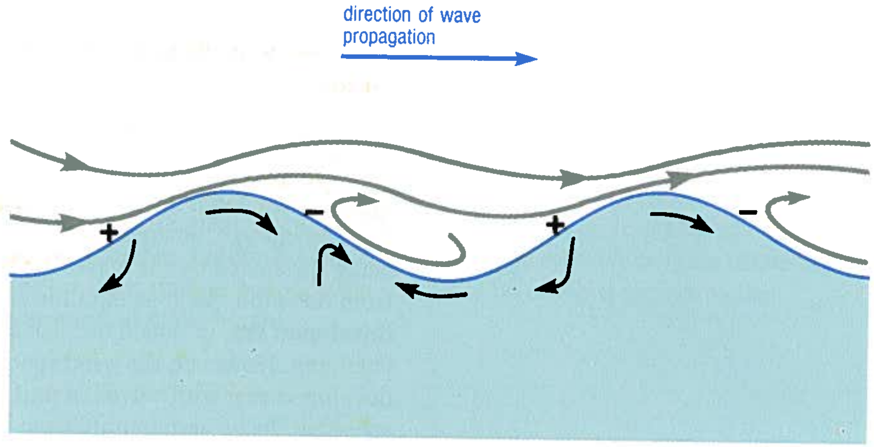


Figure 1.3 Jeffreys' 'sheltering' model of wave generation. Curved grey lines indicate air flow; shorter, black arrows show water movement. The rear face of the wave against which the wind blows experiences a higher pressure than the front face, which is sheltered from the force of the wind. Air eddies are formed in front of each wave, leading to excesses and deficiencies of pressure (shown by plus and minus signs respectively), and the pressure difference pushes the wave along.

Although Jeffreys' hypothesis fails to explain the formation of very small waves, it does seem to work if:

1 Wind speed exceeds wave speed.

2 Wind speed exceeds 1 m s^{-1}.

3 The waves are steep enough to provide a sheltering effect.

Empirically, it can be shown that the sheltering effect is at a maximum when wind speed is approximately three times the wave speed. In general, the greater the amount by which wind speed exceeds wave speed, the steeper the wave. In the open oceans, most wind-generated waves have steepness (H/L) of about 0.03 to 0.06. However, as we shall see later, wave speed in deep water is not related to wave steepness, but to wavelength – the greater the wavelength, the faster the wave travels.

QUESTION 1.2 Two waves have the same height, but differing steepness. Which of the two waves will travel the faster?

Consider the sequence of events that occurs if, after a period of calm weather, a wind starts to blow, rapidly increases to a gale, and continues to blow at constant gale force for a considerable time. No significant wave growth occurs until wind speed exceeds 1 m s^{-1}. Then, small steep waves form as the wind speed increases. Even after the wind has reached a constant gale force, the waves continue to grow with increasing rapidity until they reach a size and wavelength appropriate to a speed which corresponds to one-third of the wind speed. Beyond this point, the waves continue to grow in size, wavelength and speed, but at an ever-diminishing rate. On the face of it, one might expect that wave growth would continue until wave speed was the same as wind speed. However, in practice wave growth ceases whilst wave speed is still at some value below wind speed. This is because:

1 Some of the wind energy is transferred to the ocean surface via a tangential force, producing a surface current.

2 Some wind energy is dissipated by friction, and is converted to heat and sound.

3 Energy is lost from larger waves as a result of **white-capping**, i.e. breaking of the tip of the wave crest because it is being driven forward by the wind faster than the wave itself is travelling. Much of the energy dissipated during white-capping is converted into forward momentum of the water itself, reinforcing the surface current initiated by process 1 above.

1.1.3 THE FULLY DEVELOPED SEA

We have already seen that the size of waves in deep water is governed not only by the actual wind speed, but also by the length of time the wind has been blowing at that speed. Wave size also depends upon the unobstructed distance of sea, known as the **fetch**, over which the wind blows.

Provided the fetch is extensive enough and the wind blows at constant speed for long enough, an equilibrium is eventually reached, in which energy is being dissipated by the waves at the same rate as the waves receive energy from the wind. Such an equilibrium results in a sea state called a **fully developed sea**, in which the size and characteristics of the waves are not changing. However, the wind speed is usually variable, so the ideal fully developed sea, with waves of uniform size, rarely occurs. Variation in wind speed produces variation in wave size, so, in practice, a fully developed sea consists of a range of wave sizes known as a **wave field**. Waves coming into the area from elsewhere will also contribute to the range of wave sizes, as will interaction between waves – a process we explain in Section 1.4.2.

Oceanographers find it convenient to consider a wave field as a spectrum of wave energies (Figure 1.4). The energy contained in an individual wave is proportional to the square of the wave height (see Section 1.4).

QUESTION 1.3 Examine Figure 1.4. Does the energy contained in a wave field increase or decrease as the average frequency of the constituent waves increases?

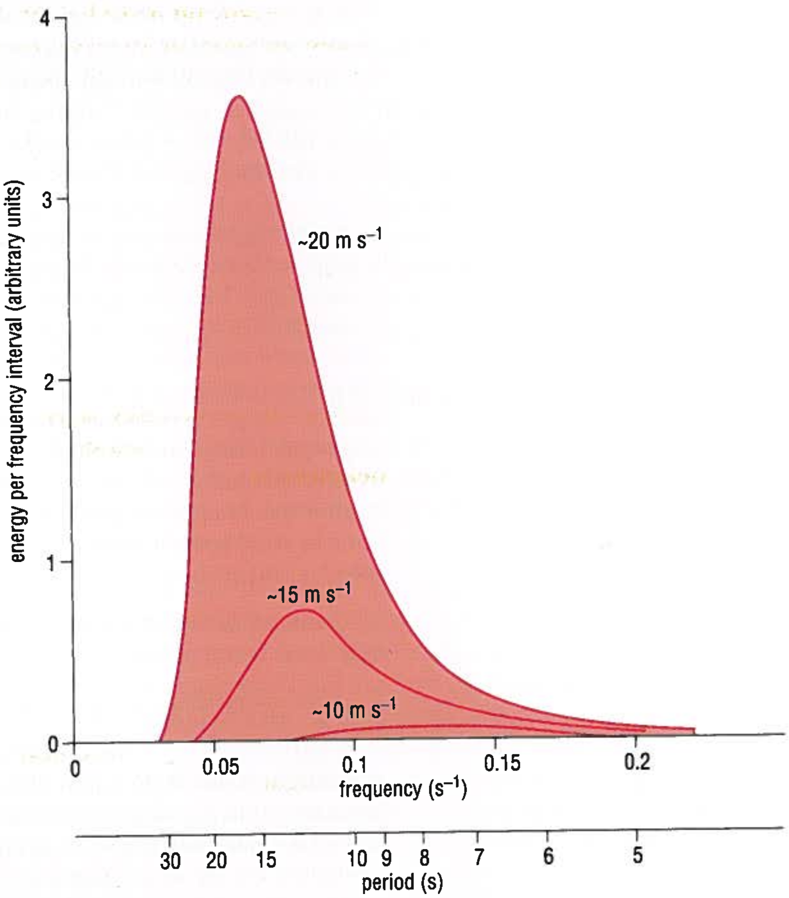


Figure 1.4 Wave energy spectra for three fully developed seas, related to wind speeds of 20, 30 and 40 knots (about 10, 15 and 20 m s^{-1} respectively). The area under each curve is a measure of the total energy in that particular wave field.

1.1.4 WAVE HEIGHT AND WAVE STEEPNESS

As was hinted in Section 1.1.3, the height of any real wave is determined by many component waves, of different frequencies and amplitudes, which move into and out of phase with, and across each other ('in phase' means that peaks and troughs coincide). In theory, if the heights and frequencies of all the contributing waves were known, it would be possible to predict the heights and frequencies of the real waves accurately. In practice, this is rarely possible. Figure 1.5 illustrates the range of wave heights occurring over a short time at one location – there is no obvious pattern to the variation of wave height.

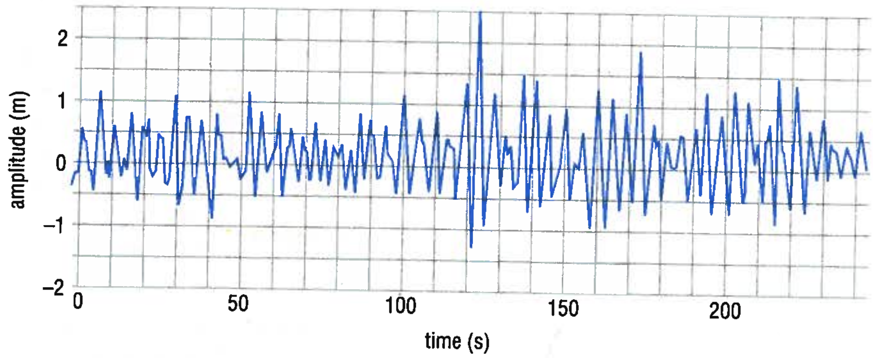


Figure 1.5 A typical wave record, i.e. a record of variation in water level (displacement from equilibrium) with time at one position.

For many applications of wave research, it is necessary to choose a single wave height which characterizes a particular sea state. Many oceanographers use the **significant wave height**, $H_{1/3}$, which is the average height of the highest one-third of all waves occurring in a particular time period. In any wave record, there will also be a maximum wave height, H_{max}. Prediction of H_{max} for a given period of time has great value in the design of structures such as flood barriers, harbour installations and drilling platforms. To build these structures with too great a margin of safety would be unnecessarily expensive, but to underestimate H_{max} could have tragic consequences. However, it is necessary to emphasize the essentially random nature of H_{max}. Although the wave $H_{max(25\ years)}$, will occur *on average* once every 25 years, this does not mean such a wave will automatically occur every 25 years – there may be periods much longer than that without one. On the other hand, two such waves might appear next week.

As wind speed increases, so $H_{1/3}$ in the fully developed sea increases. The relationship between sea state, $H_{1/3}$ and wind speed is expressed by the **Beaufort Scale** (Table 1.1, overleaf). The Beaufort Scale can be used to estimate wind speed at sea, but is valid only for waves generated within the local weather system, and assumes that there has been sufficient time for a fully developed sea to have become established (cf. Figure 1.4).

The absolute height of a wave is less important to sailors than is its steepness (H/L). As mentioned in Section 1.1.2, most wind-generated waves have a steepness in the order of 0.03 to 0.06. Waves steeper than this can present problems for shipping, but fortunately it is very rare for wave steepness to exceed 0.1. In general, wave steepness diminishes with increasing wavelength. The short choppy seas rapidly generated by local squalls are particularly unpleasant to small boats because the waves are steep, even though not particularly high. On the open ocean, very high waves can usually be ridden with little discomfort because of their relatively long wavelengths.

Table 1.1 A selection of information from the Beaufort Wind Scale.

Beaufort No.	Name	Wind speed (mean) knots	Wind speed (mean) $m\,s^{-1}$	State of the sea-surface	Significant wave height, $H_{1/3}$ (m)
0	Calm	<1	0.0–0.2	Sea like a mirror	0
1	Light air	1–3	0.3–1.5	Ripples with appearance of scales; no foam crests	0.1–0.2
2	Light breeze	4–6	1.6–3.3	Small wavelets; crests have glassy appearance but do not break	0.3–0.5
3	Gentle breeze	7–10	3.4–5.4	Large wavelets; crests begin to break; scattered white horses	0.6–1.0
4	Moderate breeze	11–16	5.5–7.9	Small waves, becoming longer; fairly frequent white horses	1.5
5	Fresh breeze	17–21	8.0–10.7	Moderate waves taking longer form; many white horses and chance of some spray	2.0
6	Strong breeze	22–27	10.8–13.8	Large waves forming; white foam crests extensive everywhere and spray probable	3.5
7	Near gale	28–33	13.9–17.1	Sea heaps up and white foam from breaking waves begins to be blown in streaks; spindrift begins to be seen	5.0
8	Gale	34–40	17.2–20.7	Moderately high waves of greater length; edges of crests break into spindrift; foam is blown in well-marked streaks	7.5
9	Strong gale	41–47	20.8–24.4	High waves; dense streaks of foam; sea begins to roll; spray may affect visibility	9.5
10	Storm	48–55	24.5–28.4	Very high waves with overhanging crests; sea-surface takes on white appearance as foam in great patches is blown in very dense streaks; rolling of sea is heavy and visibility reduced	12.0
11	Violent storm	56–64	28.5–32.7	Exceptionally high waves; sea covered with long white patches of foam; small and medium-sized ships might be lost to view behind waves for long times; visibility further reduced	15.0
12	Hurricane	>64	>32.7	Air filled with foam and spray; sea completely white with driving spray; visibility greatly reduced	>15

1.2 SURFACE WAVE THEORY

To simplify the theory of surface waves, we assume here that the wave-form is sinusoidal and can be represented by the curves shown in Figures 1.1 and 1.6. This assumption allows us to consider wave **displacement** (η) as simple harmonic motion, i.e. a sinusoidal variation in water level caused by the wave's passage. Figure 1.1 shows how the displacement varies over distance at a fixed instant in time – a 'snapshot' of the passing waves – whereas Figure 1.6 shows how wave displacement varies with time at a fixed point.

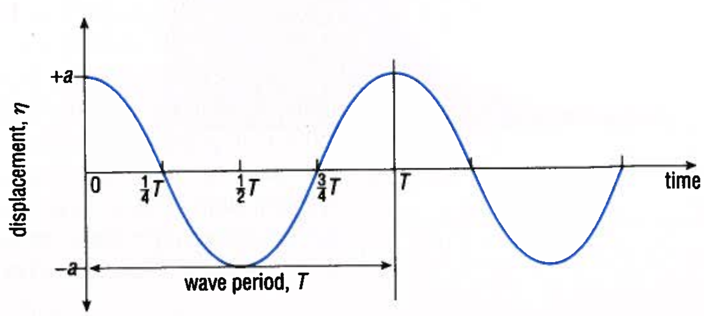


Figure 1.6 The displacement of an idealized wave at a fixed point, plotted against time. Maximum and minimum displacements are recorded in fractions of the period, T.

1.5 WAVES APPROACHING THE SHORE

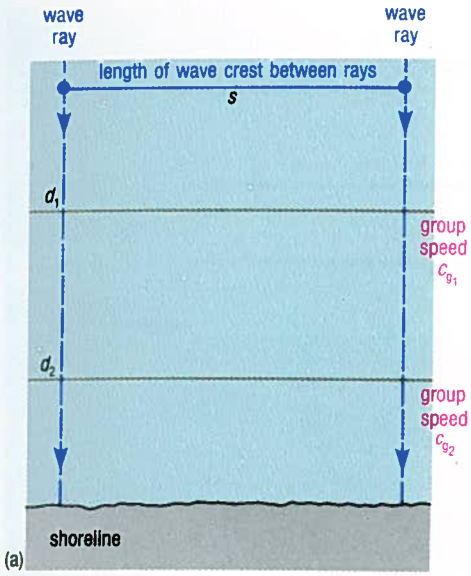


Figure 1.13 Plan view illustrating changes in the speed of waves approaching the shore. Grey lines represent wave crests at depths d_1 and d_2. Wave rays (dashed blue lines) are at right angles (i.e. normal) to the wave crests. For further explanation, see text.

It is a matter of common observation that waves coming onto a beach increase in height and steepness and eventually break. Figure 1.13 shows a length of wave crest, s, which is directly approaching a beach. As the water is shoaling, the wave crest passes a first point where the water depth (d_1) is greater than at a second point nearer the shore (where the depth is d_2). We assume that the amount of energy within this length of wave crest remains constant, the wave is not yet ready to break, and that water depth is less than 1/20 of the wavelength (i.e. Equation 1.4 applies: $c = \sqrt{gd}$). Because wave speed in shallow water is related to depth, the speed c_1 at depth d_1 is greater than the speed c_2 at depth d_2. If energy remains constant per unit length of wave crest, then

$$E_1 c_1 s = E_2 c_2 s$$

$$\text{or } \frac{E_2}{E_1} = \frac{c_1}{c_2} \qquad (1.12)$$

and because energy is proportional to the square of the wave height (Equation 1.11) then we can write

$$\frac{E_2}{E_1} = \frac{c_1}{c_2} = \frac{{H_2}^2}{{H_1}^2} \qquad (1.13)$$

Thus, both the square of the wave height and wave energy are inversely proportional to wave speed in shallow water.

This relationship is straightforward once the wave has entered shallow water. But what happens during the transition from deep to shallow water?

This is quite a difficult question, best answered by considering the highly simplified case illustrated in Figure 1.14. Imagine waves travelling shoreward over deep water (depth greater than half the wavelength). Wave speed is then governed solely by wavelength (Equation 1.3, $c = \sqrt{gL/2\pi}$. The energy is being propagated at the group speed (c_g) which is approximately half the wave speed (c), Section 1.3. As the waves move into shallower water, wave speed becomes governed by both depth and wavelength (Equation 1.2), but once the waves have moved into shallow water, where $d < L/20$, wave speed becomes governed solely by depth (Equation 1.4) and is much reduced. Remember from Section 1.3 that in shallow water group speed is equal to wave speed. The rate at which energy arrives from offshore (Figure 1.14, overleaf) must be equal to the rate at which energy moves inshore; so if the group speed in shallow water is less than half the original wave speed (and hence less than the original group speed) in deep water, the waves will show corresponding increases in height and in energy per unit area.

However, it is essential to realize that while the energy and height of individual waves will increase as they enter shallow water, *the rate of supply of wave energy* (wave power, Section 1.4.1) *must remain constant* (ignoring frictional losses).

As mentioned earlier, when waves move into shallow water, the waves begin to 'feel the bottom', the circular orbits of the water particles become flattened (Figure 1.8(c) and (d)), and wave energy will be dissipated by friction at the sea-bed, resulting in to-and-fro movement of sediments. The gentler the slope of the immediate offshore region, the sooner the incoming waves will 'feel' the bottom, the greater will be the friction with the sea-bed and the greater the energy loss before the waves finally break (see Section 1.5.2).

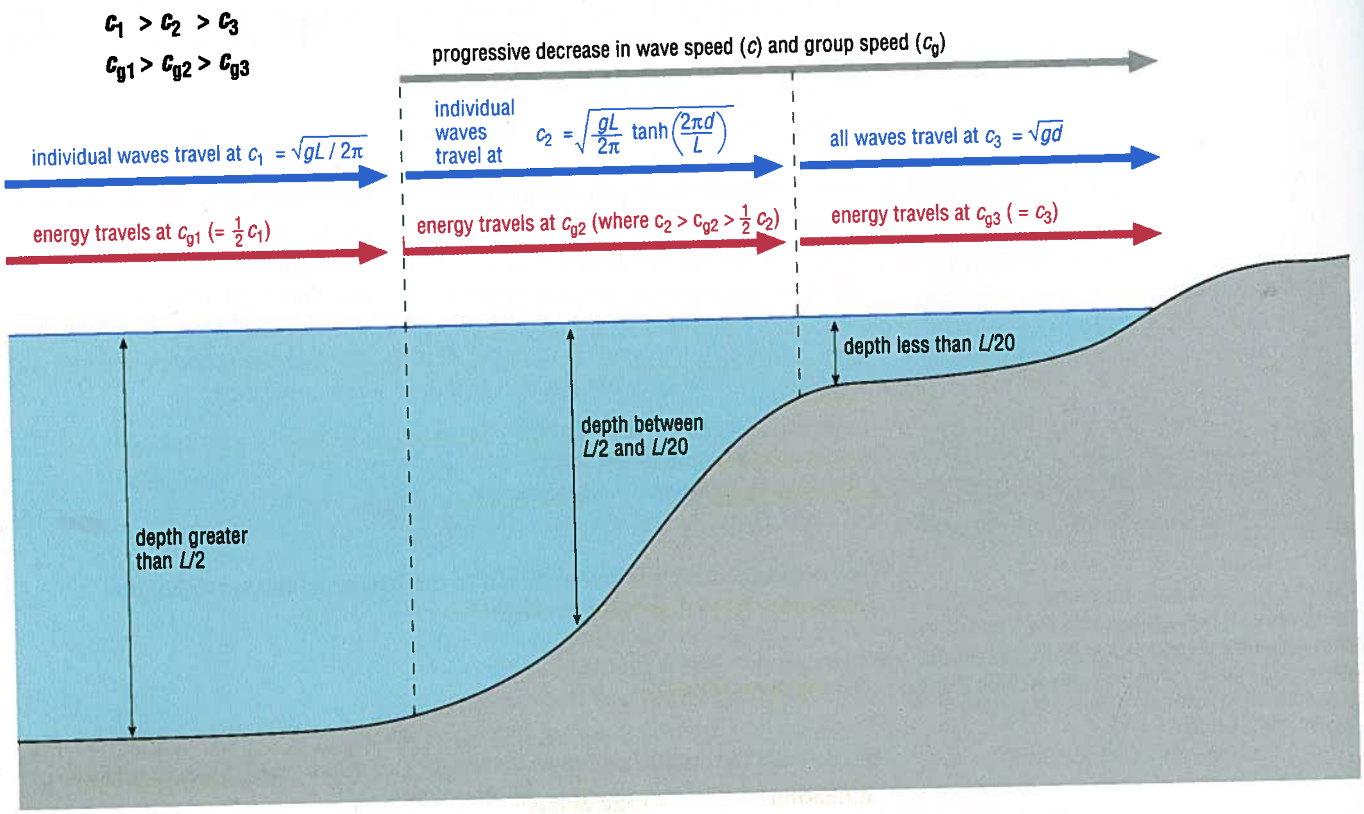


Figure 1.14 Vertical section (not to scale) illustrating changes in the relationship between wave speed and group speed, and how this affects the speed of energy propagation as waves move from deep to shallow water. The energy is being brought in from offshore at the same rate as it is being removed as the waves break at the shore (see Section 1.5.2). As group speed (c_g) in shallow water is less than in deep water, then the waves in shallow water must have more energy per unit length of wave crest, and a greater wave height than the waves in deep water.

1.5.1 WAVE REFRACTION

Figure 1.15 shows an idealized linear wave crest (length s_1, between A and B) approaching a shoreline at an angle. Because the waves are travelling in shallow water, their speed is depth-determined (Equation 1.4, $c = \sqrt{gd}$). The depth at A exceeds the depth at B, so the wave at A will travel faster than the wave at B, leading to the phenomenon known as **refraction**, which will tend to 'swing' the wave crest to an alignment parallel with the depth contours.

Can the extent of refraction be quantified?

Refraction of waves in progressively shallowing (shoaling) water can be described by a relationship equivalent to Snell's law, which describes refraction of light rays through materials of different refractive indices. Rays can be drawn perpendicular to the wave crests, and will indicate the direction of wave movement. The angle between these wave rays and lines drawn perpendicular to the depth contours can be related to wave speeds at various depths. In Figure 1.15, a wave ray approaching shoaling water at an angle θ_1, where water depth is d_1, will be at an angle θ_2 when it reaches depth d_2. Angles θ_1 and θ_2 are related to wave speed by:

$$\frac{\sin\theta_1}{\sin\theta_2} = \frac{c_1}{c_2} = \frac{\sqrt{gd_1}}{\sqrt{gd_2}} = \frac{\sqrt{d_1}}{\sqrt{d_2}} = \sqrt{\frac{d_1}{d_2}} \tag{1.14}$$

where c_1 and c_2 are the respective wave speeds at depths d_1 and d_2.

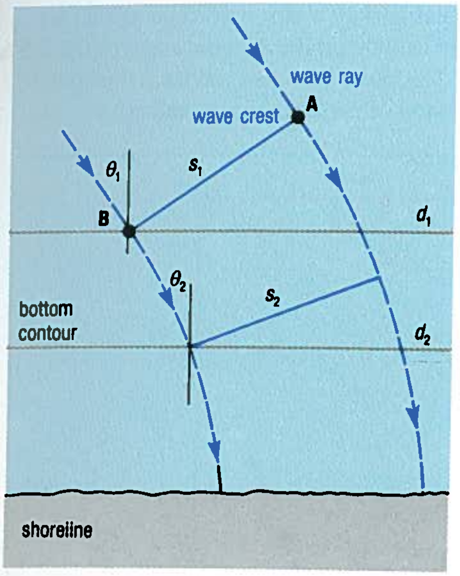


Figure 1.15 Plan view illustrating the relationship between wave approach angle (θ), water depth (d), and length of wave crest (s). The wave rays (broken blue lines), are normal to the wave crests, and are the paths followed by points on the wave crests. For further explanation, see text.

You might ask: why go to the trouble of drawing perpendiculars? Why not simply use the angles between wave crests and bottom contours?

Well, of course, one could do that, and obtain analogous relationships between the relevant angles, depths and wave speeds. However, as you will see presently, wave rays are often more useful than wave crests in determining regions that are likely to experience waves that are higher or lower than normal because of refraction.

Consider a length, s_1, of ideal wave crest, with energy per unit length E_1, which is bounded by two wave rays, as in Figure 1.15. To a first approximation, we may assume that the total energy of the wave crest between these two rays will remain constant as the wave progresses. Therefore, if the two rays converge, the same amount of energy is contained within a shorter length of wave crest, so that, for the total wave energy to remain constant, the wave height will have to increase (Equation 1.11). Conversely, if the wave rays were to diverge, then the wave height would decrease.

If, as they finally approach the shore, the two wave rays are separated by a length s_2, as in Figure 1.15, and if the wave energy is conserved, then the final wave energy must equal the initial wave energy, i.e. $E_1s_1 = E_2s_2$, or in terms of wave heights (remember Equation 1.11):

$$H_1^2s_1 = H_2^2s_2 \qquad (1.15)$$

For simplicity, s_2 in Figure 1.15 is the same length as s_1, but it is common for wave rays to converge or diverge, and in general they converge (focus) on headlands and diverge in bays (Figure 1.16).

In more complicated situations, wave refraction diagrams can be plotted for a region by using the wave of most common period and the most common direction of approach, and in this way areas in which wave rays are focused or defocused can be identified.

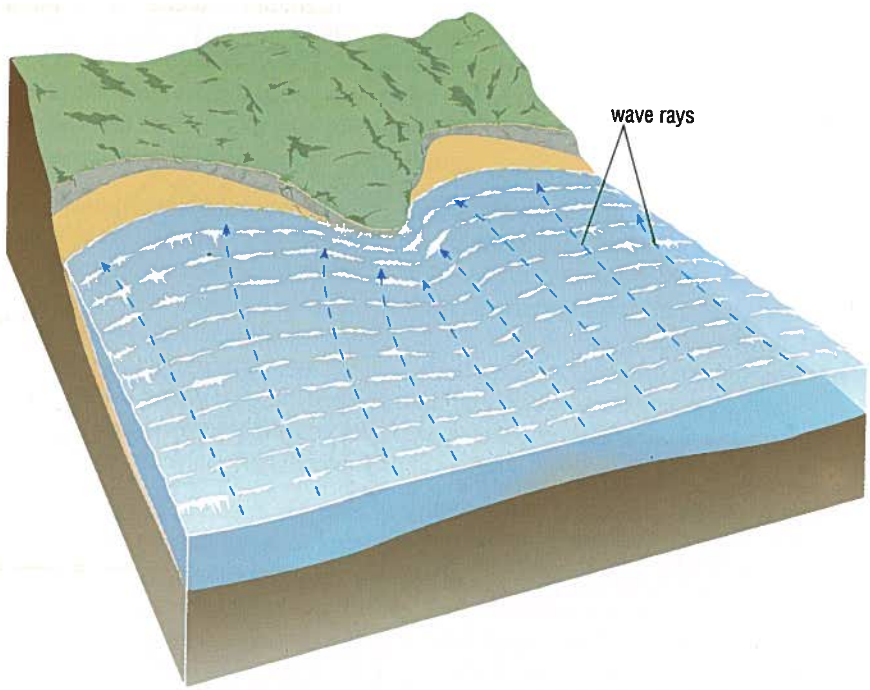


Figure 1.16 Waves are refracted and the wave rays show how wave energy is focused on headlands, where erosion is active, while deposition occurs in the bays, where the wave rays diverge and wave energy is less. Waves 'feel the bottom' and are slowed first in the shallow areas off the headland. The parts of the wave fronts that move through the deeper water leading into the bays are not slowed until they are well into the bays.

QUESTION 1.14 Figure 1.17 is a bathymetric map and storm wave refraction diagram for the Hudson River submarine canyon on the Atlantic coast of the USA. In what area covered by the refraction diagram would you advise fishermen to leave their boats to minimize the likelihood of major damage, and why?

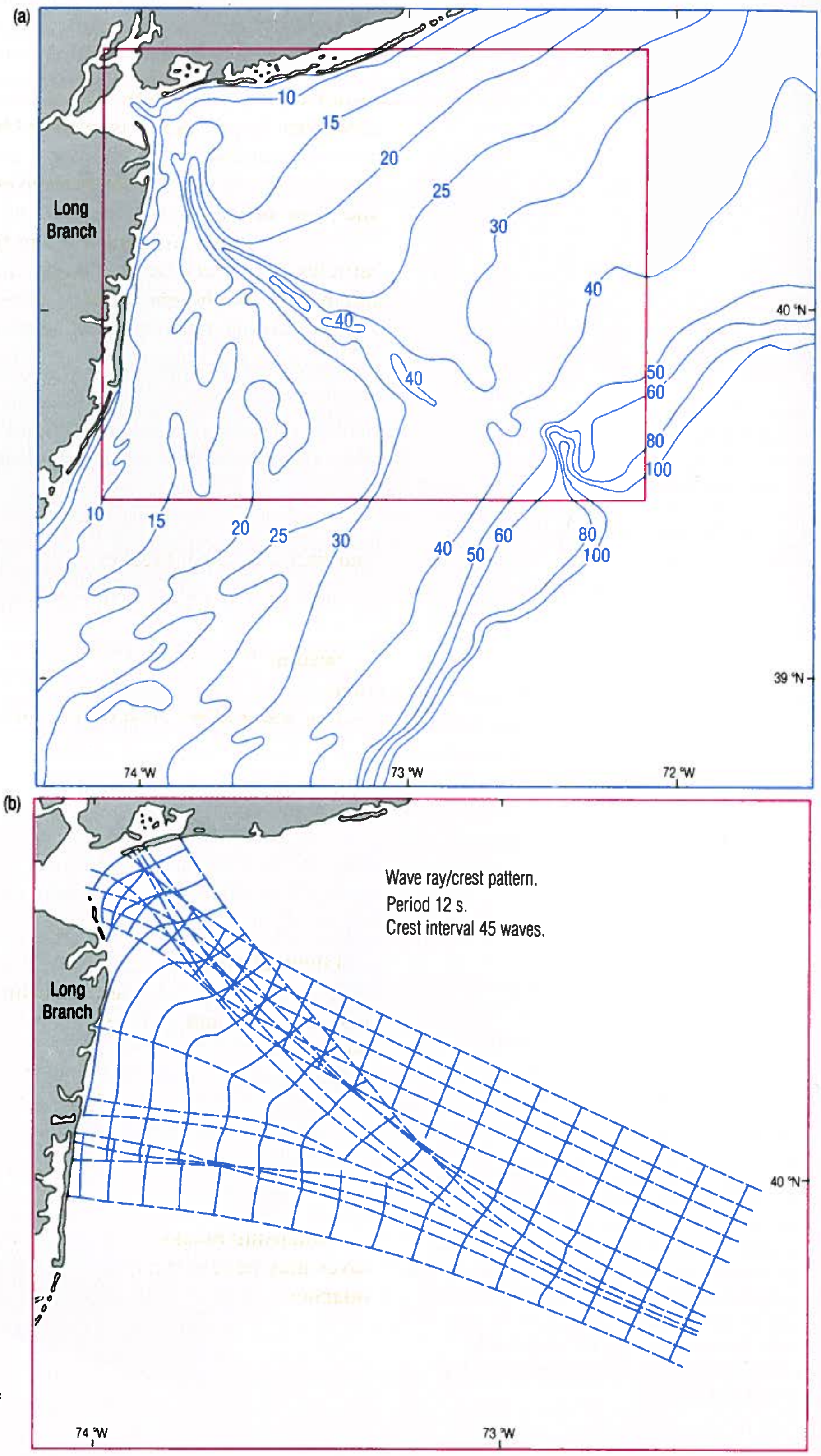


Figure 1.17 (a) Bathymetric map of the continental shelf off New York harbour at the mouth of the Hudson River on the Atlantic coast of the USA. The rectangle shows the area of map (b). The position of the Hudson Canyon can be deduced from the submarine contours (in fathoms).
(b) Pattern of wave crests (wave fronts) and wave rays off part of Long Branch beach. The wave fronts are drawn at intervals of 45 waves of period 12 s.

We can estimate increase or decrease in wave size by measuring the distances between wave rays, and applying Equation 1.15. This method is quite useful provided wave rays neither approach each other too closely nor cross over, as in these cases the waves become high, steep and unstable, and simple wave theory becomes inadequate.

1.5.2 WAVES BREAKING UPON THE SHORE

As a wave breaks upon the shore, the energy it received from the wind is dissipated. Some energy is reflected back out to sea, the amount depending upon the slope of the beach – the shallower the angle of the beach slope, the less energy is reflected. Most of the energy is dissipated as heat and sound (the 'roar' of the surf) in the final small-scale mixing of foaming water, sand and shingle. Some energy is used in fracturing large rock or mineral particles into smaller ones, and yet more may be used to move sediments and increase the height and hence the potential energy of the beach form. This last aspect depends upon the type of waves. Small gentle waves and swell tend to build up beaches, whereas storm waves tear them down (see also Chapter 5).

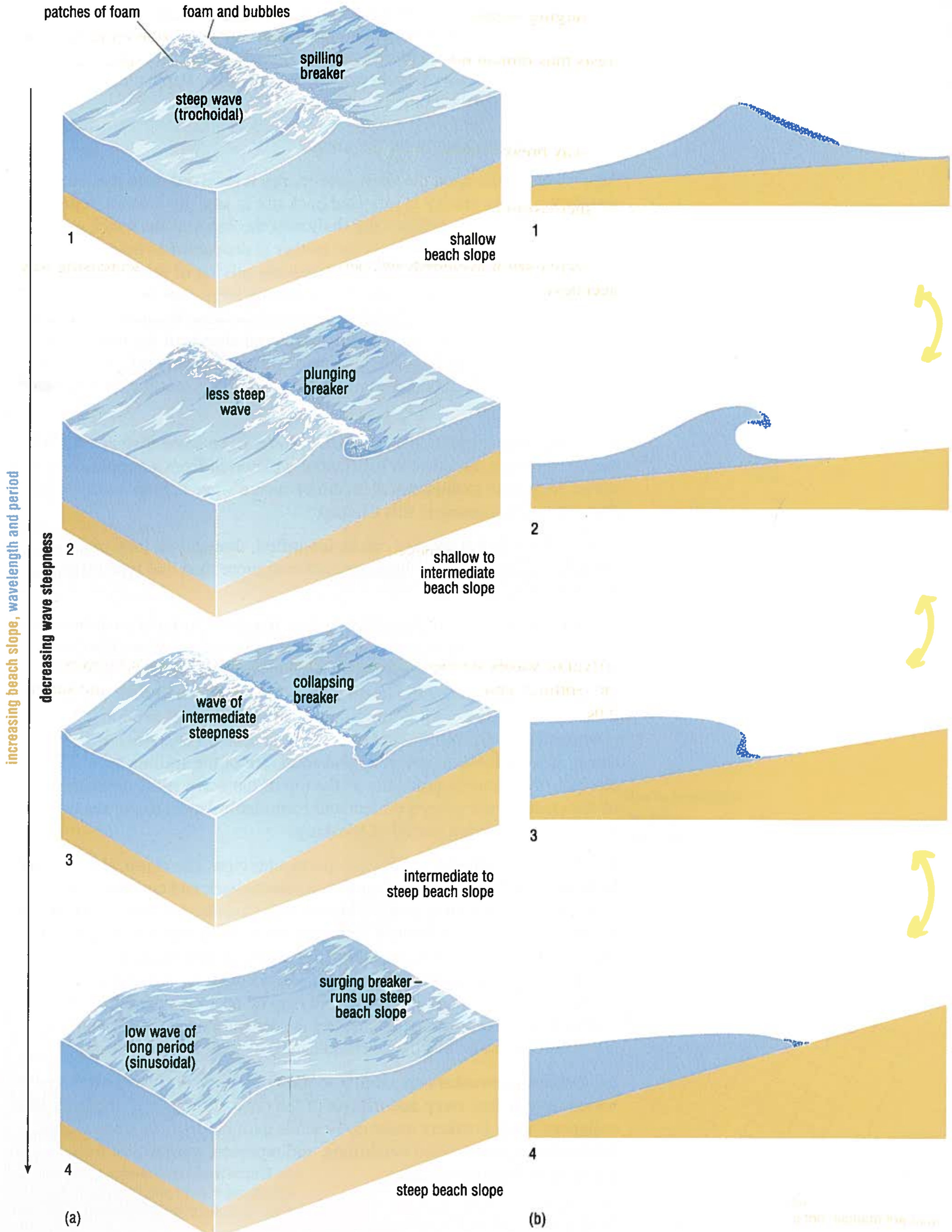
A breaking wave is a highly complex system. Even some distance before the wave breaks, its shape is substantially distorted from a simple sinusoidal wave. Hence the mathematical model of such a wave is more complicated than we have assumed in this Chapter.

Four major types of breaker can be identified, though you may often see breakers of intermediate character and/or of more than one type on the same beach at the same time.

1 Spilling breakers are characterized by foam and turbulence at the wave crest. Spilling usually starts some distance from shore and is caused when a layer of water at the crest moves forward faster than the wave as a whole. Foam eventually covers the leading face of the wave, and such waves are characteristic of a gently sloping shoreline. A tidal bore (Section 2.4.3) is an extreme form of a spilling breaker. Breakers seen on beaches *during* a storm, when the waves are steep and short, are of the spilling type. They dissipate their energy gradually as the top of the wave spills down the front of the crest, which gives a violent and formidable aspect to the sea because of the more extended period of breaking.

2 Plunging breakers are the most spectacular type. The classical form, much beloved by surf-riders, is arched, with a convex back and concave front. The crest curls over and plunges downwards with considerable force, dissipating its energy over a short distance. Plunging breakers on beaches of relatively gentle slope are usually associated with the long swells generated by distant storms. Locally generated storm waves seldom develop into plunging breakers on gently sloping beaches, but may do so on steeper ones. The energy dissipated by plunging breakers is concentrated at the *plunge point* (i.e. where the water hits the bed) and can have great erosive effect.

3 Collapsing breakers are similar to plunging breakers, except that the waves may be less steep and instead of the crest curling over, the front face collapses. Such breakers occur on beaches with moderately steep slopes, and under moderate wind conditions, and represent a transition from plunging to surging breakers.


patches of foam
foam and bubbles
spilling breaker
steep wave (trochoidal)
1
shallow beach slope
less steep wave
plunging breaker
2
shallow to intermediate beach slope
wave of intermediate steepness
collapsing breaker
3
intermediate to steep beach slope
surging breaker – runs up steep beach slope
low wave of long period (sinusoidal)
4
steep beach slope
increasing beach slope, wavelength and period
decreasing wave steepness
(a)
1
2
3
4
(b)

4 Surging breakers are found on the very steepest beaches. Surging breakers are typically formed from long, low waves, and the front faces and crests thus remain relatively unbroken as the waves slide up the beach.

Figure 1.18 illustrates the relationship between wave steepness, beach steepness and breaker type.

The way breaker shape changes from top to bottom of the picture depends upon:

1 Increasing beach slope (if considered independently from wave characteristics).

2 Increasing wavelength and period and correspondingly decreasing wave steepness, if these characteristics are considered independently of beach slope.

It is not always possible to consider 1 and 2 separately, because as you will see in later Chapters, beach slope is partly influenced by prevailing wave type and partly by the particle sizes of the beach sediments, which in turn depend upon the energy of the waves which erode, transport and deposit them.

QUESTION 1.15 If you observed plunging breakers on a beach and walked along towards a region where the beach became steeper, what different types of breaker might you expect to see?

From the descriptions, Figure 1.18, and the answer to Question 1.15, it can be seen that the four types of breaker form a continuous series. The spilling breaker, characteristic of shallow beaches and steep waves (i.e. with short periods and large amplitudes), forms one end of the series. At the other end of the series is the surging breaker, characteristic of steep beaches and of waves with long periods and small amplitudes. For a given beach, the arrival of waves steeper than usual will tend to give a type of breaker nearer the 'spilling' end of the series, whereas calmer weather favours the surging type. The dynamics of collapsing (3) and surging (4) breakers are affected by bottom slope more than those of spilling (1) and plunging (2) breakers. Spilling and plunging breakers can also occur in deep water, partly because the sea-bed is far below and does not affect wave dynamics. Collapsing and surging breakers do not occur in deep water.

(c)

Figure 1.18 (a) The four types of breaker seen in perspective view from top to bottom (1–4): spilling, plunging, collapsing, surging. The vertical arrow shows their relationships to beach slope, wave period, length and steepness.
(b) Cross-sections through the four breaker types.
(c) Photograph of a breaker, part spilling, part plunging. See text for further discussion.

IMPORTANT: When examining Figure 1.18, you need to be aware that the four types of breaker illustrated are just stages in a continuous spectrum; changes from one to another are gradual, not instantaneous.