WAVE PREDICTION AND FORECASTING

See **WAVE GENERATION BY WIND**

WAVES ON BEACHES

R. A. Holman, Oregon State University,
Corvallis, OR, USA

doi:10.1006/rwos.2001.0131

Introduction

Wave motions are one of the most familiar of oceanographic phenomena. The waves that we see on beaches were originally generated by ocean winds and storms, sometimes at long distances from their final destination. In fact, groups of waves, generated by large storms, have been tracked from the Southern Ocean near Australia all the way to Alaska.

Open ocean waves can be thought of as simple sinusoids that are superimposed to yield a realistic sea. Waves entering the nearshore, called incident waves, can have wave periods (T, the time between consecutive passages of wave crests) ranging from 2 to 20 s, with 10 s a typical value. Wave heights (H, the vertical distance from the trough to peak of a wave) can exceed 10 m, but are typically 1 m, representing an energy density, of $1250\,\mathrm{J\,m^{-2}}$ (ρ is the density of sea water, g is the acceleration of gravity) and a flux of power impinging on the coast of about 10 kW per meter of coastline. Although this is a substantial amount of power, it is not enough to make broad commercial exploitation of wave power economical at the time of writing.

Of interest in this section are the dynamics of waves once they progress into the shallow beach environment such that the ocean bottom begins to restrict the water motions. Most people are familiar with refraction (the turning of waves toward the beach), wave breaking, and swash (the back and forth motion of the water's edge), but are less familiar with the other types of fluid motion that are generated near the beach.

Figure 1 illustrates schematically the evolution of ocean wave energy as it moves from deep water (top of the figure) through progressively shallower water toward the beach (bottom of the figure). Offshore, most energy lies in waves of roughly 10 s period (middle of the figure). However, the processes of shoaling distribute that energy to both higher frequencies (right half of the figure) and lower frequencies (left half) including mean flows. In general, these processes can be distinguished as those that occur offshore of the surf zone (where waves become overly steep and break) and those that occur within the surf zone.

The axes of **Figure 1**, cross-shore position and frequency, are two of several variables that can be used to structure a discussion of near-shore fluid

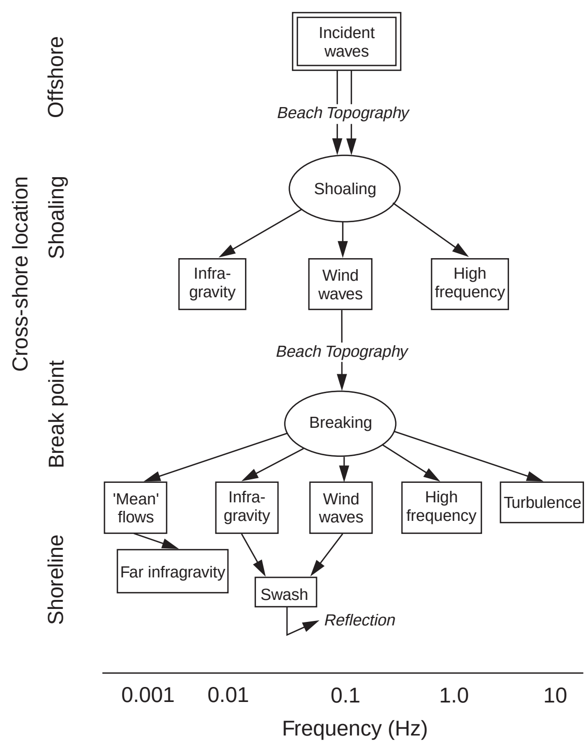

Figure 1 Schematic of important near-shore processes showing how the incident wave energy that drives the system evolves as the waves progress from offshore to the shoreline (top to bottom of figure). Wave evolution is grouped into processes occurring seaward of the breakpoint (labeled 'shoaling') and those within the surf zone (denoted 'breaking'). In both cases, energy is spread to lower (left) and higher (right) frequencies. The beach topography provides the bottom boundary condition for flow, so is important to wave processes. In turn, the waves move sediment, slowly changing the topography. Wind and tides may be important in some settings, but are not shown here.

Table 1 Kinematic relationships of linear waves[a]

Parameter	*Shallow water*[b]	*General expression*	*Deep water*[c]
Surface elevation	$\eta = \frac{H}{2}\cos(kX - \sigma t)$	$\eta = \frac{H}{2}\cos(kX - \sigma t)$	$\eta = \frac{H}{2}\cos(kX - \sigma t)$
Energy density	$E = \frac{1}{8}\rho g H^2$	$E = \frac{1}{8}\rho g H^2$	$E = \frac{1}{8}\rho g H^2$
Wave length	$L = T\sqrt{(gh)}$	$L = \frac{gT^2}{2\pi}\tanh(kh)$	$L_0 = \frac{gT^2}{2\pi}$
Phase velocity	$c = \sqrt{(gh)}$	$c = \frac{gT}{2\pi}\tanh(kh)$	$c_0 = \frac{gT}{2\pi}$
Group velocity	$c_g = c = \sqrt{(gh)}$	$c_g = cn = c\left[\frac{1}{2} + \frac{kh}{\sinh(2kh)}\right]$	$c_g = \frac{c}{2} = \frac{gT}{4\pi}$
Wave power, $P = Ec_g$	$P = \frac{1}{8}\rho g H^2 \sqrt{(gh)}$	$\frac{\rho g H^2}{8}\frac{gT\tanh(kh)}{2\pi}\left[\frac{1}{2} + \frac{kh}{\sinh(2kh)}\right]$	$P = \frac{1}{8}\rho g H^2 \frac{gT}{4\pi}$
Wave momentum flux	$S_{xx} = \frac{3}{2}E$	$S_{xx} = E\left[\frac{2kh}{\sinh(2kh)} + \frac{1}{2}\right]$	$S_{xx} = \frac{1}{2}E$
Horizontal velocity magnitude	$u = \frac{H}{2}\sqrt{\left(\frac{g}{h}\right)}$	$u = \frac{\pi H}{T}\frac{\cosh(k(z+h))}{\sinh(kh)}$	$u = \frac{\pi H}{T}e^{kz}$

[a] H is the peak to trough wave height and is twice the amplitude, a; k is the wave number $= 2\pi/L$; X is distance in the direction of propagation; σ is the frequency $= 2\pi/T$ where T is the wave period; ρ is the density of water; g is the acceleration of gravity; z is the depth below the surface within the water column of total depth h. Units of each quantity depend on units used in each equation.
[b] $h < L_0/20$. h is the water depth; L_0 is the deep water wavelength.
[c] $h < L_0/2$.

dynamics. Other important distinctions that will be made include whether the incident waves are monochromatic (single frequency) versus random (including a range of frequencies), depth-averaged versus depth-dependent, longshore uniform (requiring consideration of only one horizontal dimension, 1HD) versus long shore variable (2HD), and linear versus nonlinear.

The Dynamics of Incident Waves

Much of the early progress in understanding nearshore waves was based on the examination of a monochromatic wave train, propagating onto a long shore uniform beach (1HD). Many observable properties can be explained in terms of a few principles including conservation of wave crests, of momentum, and of energy. Most dynamics are depth-averaged.

Table 1 lists a number of properties of monochromatic ocean waves, in the linear limit of infinitesimal wave amplitude (known as linear, or Airy wave theory). The general expressions (center column) contain complicated forms that can be substantially simplified for both shallow (depths less than 1/20 of the deep water wavelength, L_0) and deep (depths greater than 1/2 L_0) water limits.

The speed of wave propagation is known as the celerity or phase speed, c, to distinguish it from the velocity of the actual water particles. In deep water, c depends only on the wave period (independent of depth). However, as the wave enters shallower water, the wavelength decreases and the phase speed becomes slower (contrary to common belief, this is not a result of bottom friction). An interesting consequence of the slowing is wave refraction, the turning of waves toward the coast. For a wave approaching the coast at any angle, the end in shallower water will always progress more slowly than the deeper end. By propagating faster, the deeper end will begin to catch up to the shallow end, effectively turning the wave toward the beach (refraction). In shallow water, the general expression for celerity, $c = (gh)^{1/2}$, depends only on depth so that waves of all periods propagate at the same speed.

The energy density of Airy waves (energy per unit area) is the sum of kinetic and potential energy components and depends only on the square of the wave height (**Table 1**). Perhaps of more interest is the rate at which this energy is propagated by the wave train, known as the wave power, P, or wave energy flux. In deep water, wave energy progresses at half the speed of wave phase (individual wave crests will out-run the energy packet), whereas in shallow water energy travels at the same speed as wave phase and the flux depends only on $H^2h^{1/2}$. Offshore of the surf zone, wave energy is conserved since there is no breaking dissipation and energy loss through bottom friction has been shown to be negligible except over very wide flat seas. Thus, as the depth, h, decreases, the wave height, H, must increase to conserve $H^2h^{1/2}$. This is a phenomenon familiar to beach-goers as the looming up of a wave just before breaking.

The combined result of shoaling is reduced wavelength and increased wave height, hence waves that become increasingly steep and may become unstable and break. One criterion for breaking is that the increasing water particle velocities (u in **Table 1**) exceed the decreasing wave phase speed such that the water leaps ahead of the wave in a curling or plunging breaker. From the relationships in **Table 1**, it can be found that this occurs. when γ, the wave height to depth ratio ($\gamma = H/h$) exceeds a value of 2. Of course, for waves that have steepened to the point of breaking, the approximations of infinitesimal waves, inherent in Airy wave theory, are badly violated. However, observations show that γ does reach a limiting value of approximately 1 for monochromatic waves and 0.4 for a random wave field. As waves continue to break across the surf zone, the wave height decreases in a way that γ is approximately maintained, and the wave field is said to be saturated (cannot get any larger).

The above-saturation condition implies that wave heights will be zero at the shoreline and there will be no swash, in contradiction to common observation. Instead, it can be shown that very small amplitude waves, incident on a sloping beach, will not break unless their shoreline amplitude, a_s, exceeds a value determined by

$$\frac{\sigma^2 a_s}{g\beta^2} \leq \kappa \qquad [1]$$

where κ is an O(1) constant. For larger amplitudes, the wave amplitude at any cross-shore position is the sum of a standing wave contribution of this maximum value plus a dissipative residual that obeys the saturation relationship.

The ratio of terms on the left-hand side of eqn. [1] is important to a wide range of nearshore phenomena and is often re-written as the Iribarren number,

$$\xi_0 = \frac{\beta}{(H_s/L_0)^{1/2}} \qquad [2]$$

where the measure of wave amplitude is replaced by the offshore significant wave height,[1] H_s. This form clarifies the importance of beach steepness, made dynamically important by comparing it to the wave steepness, H_s/L_0. For very large values of ξ_0, the beach acts as a wall and is reflective to incident waves (the non-breaking case from eqn. [1]). For smaller values, the presence of the sloping beach takes on increasing importance as the waves begin to break as the plunging breakers that surfers like, where water is thrown ahead of the wave and the advancing crest resembles a tube. Still smaller beach steepnesses (and ξ_0) are associated with spilling breakers in which a volume of frothy turbulence is pushed along with the advancing wave front.

Radiation Stress: the Forcing of Mean Flows and Set-up

The above discussion is based on the assumption of linearity, strictly true only for waves of infinitesimal amplitude. Because the dynamics are linear, energy in a wave of some particular period, say 10 s, will always be at that same period. In fact, once wave amplitude is no longer negligible, there are a number of nonlinear interactions that may transfer energy to other frequencies, for example to drive currents (zero frequency).

Nonlinear terms describe the action of a wave motion on itself and arise in the momentum equation from the advective terms, $u \cdot \nabla u$, and from the integrated effect of the pressure term. For waves, the time-averaged effect of these terms can easily be calculated and expressed in terms of the radiation stress, S, defined as the excess momentum flux due to the presence of waves. Since a rate of change of momentum is the equivalent of a force by Newton's second law, radiation stress allows us to understand the time-averaged force exerted by waves on the water column through which they propagate. A spatial gradient in radiation stress, for instance a larger flux of momentum entering a particular location than exiting, would then force a current.

[1] Although the peak to trough vertical distance for monochromatic waves is a unique and hence sensible measure of wave height, for random waves this scale is a statistical quantity, representing a distribution. A single measure, often chosen to represent the random wave field, is the significant wave height, H_s, defined as the average height of the largest one-third of the waves. This statistic was chosen historically as best representing the value that would be visually estimated by a semitrained observer. It is usually calculated as four times the standard deviation of the sea surface times series.

Radiation stress is a tensor such that S_{ij} is the flux of i-directed momentum in the j-direction. For waves in shallow water, approaching the coast at an angle θ, the components of the radiation stress tensor are

$$S = \begin{bmatrix} S_{xx} & S_{xy} \\ S_{yx} & S_{yy} \end{bmatrix} = E\begin{bmatrix} (\cos^2\theta + 1/2) & \cos\theta\sin\theta \\ \cos\theta\sin\theta & (\sin^2\theta + 1/2) \end{bmatrix} \quad [3]$$

where x is the cross-shore distance measured positive to seaward from the shoreline and y is the long-shore distance measured in a right-hand coordinate system with z positive upward from the still water level.

For a wave propagating straight toward the beach ($\theta = 0$) $S_{xx} = 3/2\,E$ is the shoreward flux of shoreward-directed momentum. The increase of wave height (hence energy, E) associated with shoaling outside the surf zone must be accompanied by an increasing shoreward flux of momentum (radiation stress). This gradient, in turn, provides an offshore thrust, pushing water away from the break point and yielding a lowering of mean sea level called set-down. As the waves start to break and decrease in height through the surf zone, the decreasing radiation stress pushes water against the shore until an opposing pressure gradient balances the radiation stress gradient. The resulting set-up at the shoreline, $\bar{\eta}_{\max}$, a contributor to coastal erosion and flooding, is found to depend on the offshore significant wave height, H_s, as

$$\bar{\eta}_{\max} = KH_s\xi_0 \quad [4]$$

where K is found empirically to be 0.45.

If waves approach the beach at an angle, they also carry with them a shoreward flux of longshore-directed momentum, S_{yx}. Cross-shore gradients in this quantity, due to the breaking of waves in the surf zone, provide a net long shore force that accelerates a long shore current, $\bar{V}$ along the beach until the forcing just balances bottom friction. If the cross-shore structure of $\bar{V}$ is solved for, a discontinuity is evident at the seaward limit of the surf zone, where the radiation stress forcing jumps from zero (seaward of the break point) to a large value (where the wave just begin to break).

This discontinuity is an artifact of the fact that every wave breaks at exactly the same location for an assumed monochromatic wave forcing, and must be artificially smoothed by an assumed horizontal mixing for this case. However, a natural random wave field consists of an ensemble of waves with (for linear waves) a Rayleigh distribution of heights. Depth-limited breaking of such a wave field will be spread over a region from offshore, where a few largest waves break, to onshore where the smallest waves finally begin to dissipate. The spatially distributed nature of these contributions to the average radiation stress provides a natural smoothing, often obviating the need for additional horizontal smoothing.

Nonlinear Incident Waves

The above discussion dwelt on the nonlinear transfer of energy from incident waves to mean flows. Nonlinearities will also transfer energy to higher frequencies, yielding a transformation of incident wave shape from sinusoidal to peaky and skewed forms. The Ursell number, $(H/L)(L/h)^3$, measures the strength of the nonlinearity. For monochromatic incident waves, this evolution was often modeled in terms of an ordered Stokes expansion of the wave form to produce a series of harmonics (multiples of the incident wave frequency) that are locked to the incident wave. For waves with Ursell number of O(1), propagating in depths that are not large compared to the wave height, higher order theories must be used to model the finite amplitude dynamics. For a random sea under such theories, the total evolution of the spectrum must be found by summing the spectral evolution equations for all possible Fourier pairs (in other words, all frequencies in the sea can and will interact with all other frequencies). Such approaches are very successful in predicting the evolving shape and nonlinear statistics (important for driving sediment transport) for natural random wave fields outside the surf zone.

Vertically Dependent Processes

Depth-independent models are successful in reproducing many nearshore fluid processes but cannot explain several important phenomena, for example undertow, offshore-directed currents that exist in the lower part of the water column under breaking waves. The primary cause of depth dependence arises from wave-breaking processes. When waves break, the organized orbital motions break down, either through the plunge of a curling jet of water thrown ahead of the advancing wave or as a turbulent foamy mass (called a roller) carried on the advancing crest. Both processes originate at the surface but drive turbulence and bubbles into the upper part of the water column.

The transfer of momentum from wave motions to mean currents described by radiation stress gradients above does not account for the existence of an intermediate repository, the active turbulence of the roller, that decays slowly as it is carried with the progressing wave. This time delay causes a shift of the forcing of longshore currents, such that a current jet will occur landward of the location expected from study of the breaking locations of incident waves.

The other consequence of the vertical dependence of the momentum transfer is that the shoreward thrust provided by wave breaking is concentrated near the surface. Set-up, the upward slope of sea level against the shore, will balance the depth averaged wave forcing. However, due to the vertical structure of the forcing, shoreward flows are driven near the top of the water column and a balancing return flow, the undertow, occurs in the lower water column. Undertow strengths can reach $1\,\mathrm{m\,s^{-1}}$.

2HD Flows – Circulation

All of the previous discussion was based on the assumption that all processes were long-shore uniform (1 HD) so that no long-shore gradients existed. It is rare in nature to have perfect long shore uniformity. Most commonly, some variability (often strong) exists in the underlying bathymetry. This can lead to refractive focusing (the concentration of wave energy by refraction of waves onto a shallower area) and the forcing of long-shore gradients in wave height, hence of setup. Since setup is simply a pressure head, long-shore gradients will drive long-shore currents toward low points where the converging water will turn seaward in a jet called a rip current.

It is possible to develop long-shore gradients in wave height (hence rips) in the absence of long-shore variations in bathymetry. Interactions between two elements of the wave field (either two incident wave trains from different directions or an incident wave and an edge wave, defined below) can force rip currents if the interacting trains always occur with a fixed relative phase.

Infragravity Waves and Edge Waves

There is a further, very important consequence of the fact that natural wave fields are not monochromatic, but instead are random. For random waves, wave height is no longer constant but varies from wave to wave. Usually these variations are in the form of groups of five to eight waves, with heights gradually increasing then decreasing again. This observation is known as surf beat and is particularly familiar to surfers. A consequence of these slow variations is that the radiation stress of the waves is no longer constant, but also fluctuates with wave group timescales and forces flows (and waves) in the near-shore with corresponding wave periods. These waves have periods of 30–300 s and are called infragravity waves, in analogy to infrared light being lower frequency than its visible light counterpart.

The direct forcing of infragravity motions described above can be thought of as a time-varying setup, with the largest waves in a group forcing shoreward flows that pile up in setup, followed by seaward flow as the setup gradients dominate over the weaker forcing of the small waves. If the modulations of the incident wave group are long shore uniform, this result of this setup disturbance will simply be a free (but low frequency) wave motion that propagates out to sea. However, in the normal case of wave groups with longshore (as well as time) variability, we can think of the response by tracing rays as the setup disturbance tries to propagate away. Rays that travel offshore at an angle to the beach will refract away from the beach normal (essentially the opposite of incident wave refraction, discussed earlier). For rays starting at a sufficiently steep angle to the normal, refraction can completely turn the rays such that they re-approach and reflect from the shore in a repeating way and the energy is trapped within the shallow region of the beach. These trapped motions are called edge waves because the wave motions are trapped in the near-shore wave guide. (Any region wherein wave celerity is a minimum can similarly trap energy by refraction and is known as a wave guide. The deep ocean sound channel is a well-known example and allows propagation of trapped acoustic energy across entire ocean basins.) Motions that do not completely refract and thus are lost to the wave guide are called leaky modes.

The requirement that wave rays start at a sufficiently steep angle to be trapped by refraction can be expressed in terms of the long-shore component of wavenumber, k_y. For large k_y (waves with a large angle to the normal), rays will be trapped in edge waves whereas small k_y motions will be leaky modes. The cutoff between these is σ^2/g.

In the same sense that waves that slosh in a bathtub occur as a discrete set of modes that exactly fit into the tub, edge waves occur in a set of modes that exactly fit between reflection at the shoreline and an exponentially decaying tail offshore. The detailed form of the waves depends on the details of the bathymetry causing the refraction. However, for the example of a plane beach of slope β $(h = x \tan \beta)$, the cross-shore forms of the

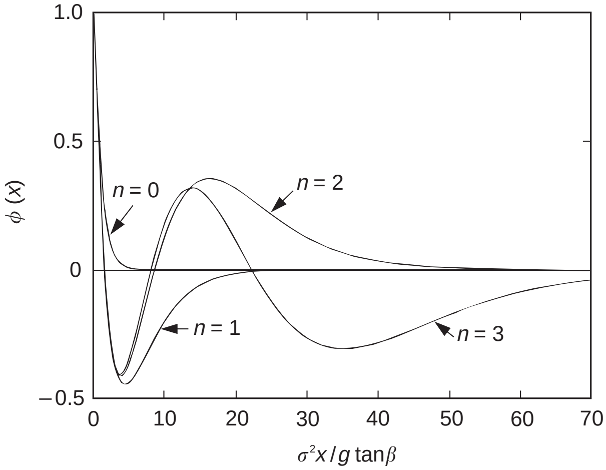

Figure 2 Cross-shore structure of edge waves. Only the lowest four mode numbers of the larger set are shown. The mode number, n, describes the number of zero crossings of the modes. So, for example, a mode 1 edge wave always has a low offshore, opposite a shoreline high, and visa versa. Edge waves propagate along the beach.

lowest four modes (mode numbers, $n = 0, 1, 2, 3$), are shown in **Figure 2** and are given in **Table 2**.

The existence of edge waves as a resonant mode of wave energy transmission in the near-shore has several impacts. First, the dispersion relation provides a selection for particular scales. For example, **Figure 3** shows a spectrum of infragravity wave energy collected at Duck, North Carolina, USA. The concentration of energy into very clear, preferred scales is striking and has led to suggestions that edge waves may be responsible for the generation of sand bars with corresponding scales. Second, because edge-wave energy is trapped in the near-shore, it can build to substantial levels even in the presence of weak, incremental forcing. Moreover, because edge-wave energy is large at the shoreline where the incident waves have decayed to their minimum

Table 2 Edge wave kinematics

$k_y < \frac{\sigma^2}{g}$, leaky modes	
$k_y > \frac{\sigma^2}{g}$, edge waves	
Velocity potential	$\Phi = \frac{ag}{\sigma}\phi(x)\cos(ky - \sigma t)$
Sea surface elevation	$\eta = -a\phi(x)\sin(ky - \sigma t)$
Cross-shore velocity	$u = \frac{ag}{\sigma}\frac{\partial\phi(x)}{\partial x}\cos(ky - \sigma t)$
Long-shore velocity	$v = -\frac{a\sigma}{\sin[(2n+1)\beta]}\phi(x)\sin(ky - \sigma t)$
Dispersion relationship	$L = \frac{gT^2}{2\pi}\sin[(2n+1)\beta]$
Cross-shore shape functions	n — $\phi(x)$
	0 — $1 \cdot e^{-kx}$
	1 — $(1 - 2kx) \cdot e^{-kx}$
	2 — $(1 - 4kx + 2k^2x^2) \cdot e^{-kx}$
	3 — $(1 - 6kx + 6k^2x^2 - 4/3k^3x^3) \cdot e^{-kx}$

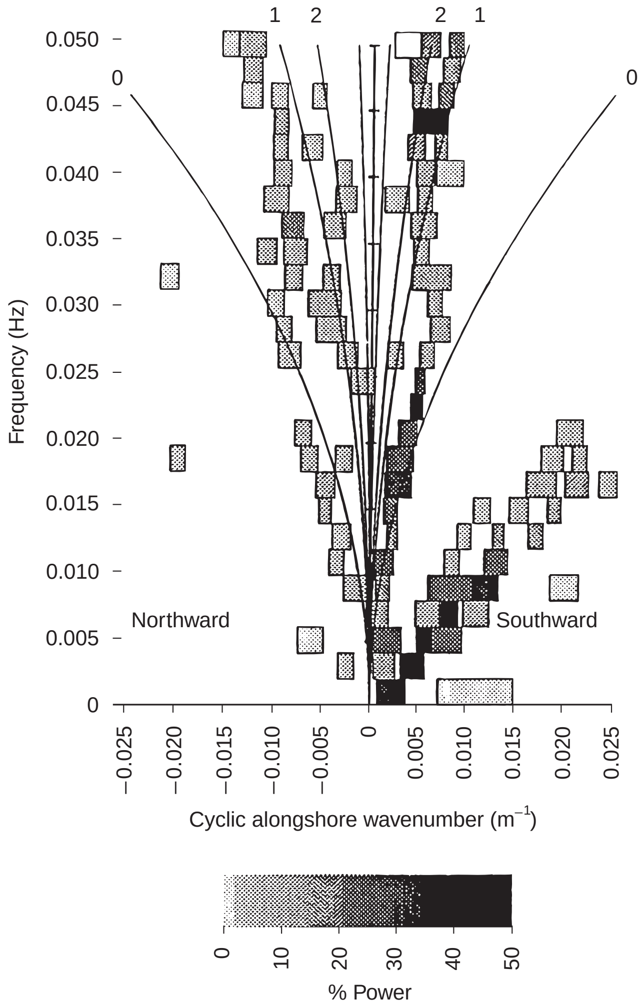

Figure 3 The spectrum of low frequency wave motions, as measured by current meters sampling the long-shore component of velocity, from Duck, North Carolina, USA. The vertical axis corresponds to frequency and the horizontal axis to alongshore component of wavenumber. Positive wavenumbers describe wave motions propagating along the beach to the south. Surprisingly, wave energy is not spread broadly in frequency–wavenumber space, but concentrates on specific ridges. Black lines, indicating the theoretical dispersion lines for edge waves (mode numbers marked at figure top), provide a good match to much of the data at low frequencies with some offset at higher frequency associated with Doppler shifting by the long-shore current. The concentration of low-frequency energy angling up to the right corresponds to shear waves. (After Oltman-Shay J and Guza RT (1987) Infragravity edge wave observations on two California beaches. *Journal of Physical Oceanography* 17(5): 644–663.)

Table 3 Magnitude of infragravity waves on different beach types

Location	ξ_0	m^a
New South Wales, Australia	0.40	0.185 ± 0.157
Duck, NC, USA	0.43	0.237 ± 0.057
Martinique Beach, Canada	0.57	0.323 ± 0.073
Torrey Pines, CA, USA	0.60	0.554 ± 0.100
Santa Barbara, CA, USA	0.83	0.459 ± 0.089

[a]Least squares regression slope between the significant swash height (computed from the infragravity band energy only) and the offshore significant wave height.

Reproduced from Howd PA, Oltman-Shay J, and Holman RA (1991) Wave variance partitioning in the trough of a barred beach. *Journal of Geophysical Research* 96 (C7), 12781–12795.)

due to breaking, edge waves may feasibly be the dominant fluid-forcing pattern on near-shore sediments in these regions.

The magnitudes of infragravity energy (including edge waves and leaky modes) have been found to depend on the relative beach steepness as expressed by the Iribarren number (eqn. [2]). For steep beaches (high ξ_0), very little infragravity energy can be generated and the beaches are termed reflective due to the high reflection coefficient for the incident waves. However, for low-sloping beaches (small ξ_0), infragravity energy can be dominant, especially compared to the highly dissipated incident waves. On the Oregon Coast of the USA, for example, swash spectra have been analyzed in which 99% of the variation is at infragravity timescales (making beachcombing an energetic activity). **Table 3** lists five representative beach locations, with mean values of ξ_0 and of m, the linear regression slope between the measured significant swash magnitude,[2] R_s, in the infragravity band, and the offshore significant wave height, H_s.

Shear Waves

Up until the mid-1980s long shore currents were viewed as mean flows whose dynamics were readily described as in the above sections. However, field data from the Field Research Facility in Duck, North Carolina, provided surprising evidence that as long-shore currents accelerated on a beach with a well-developed sand bar, the resulting current was not steady but instead developed slow fluctuations in strength and a meandering pattern in space. Typical wave periods of these wave are hundreds of seconds and long-shore wavelengths are just hundreds of meters (**Figure 3**). These very low frequencies are called far infragravity waves, in analogy to the relationship of far infrared to infrared optical frequencies. However, the wavelengths are several orders of magnitude shorter than that which would be expected for gravity waves (e.g., edge waves or leaky modes) of similar periods.

These meanders have been named shear waves and arise due to an instability of strong currents, similar to the instability of a rising column of smoke. The name comes from the dependence of the dynamics (described briefly below) on the shear of the long-shore current (the cross-shore gradient of the long-shore current). A jet-like current with large shear, such as might develop on a barred beach where the wave forcing is concentrated over and near the bar, can develop strong shear waves. In contrast, for a broad, featureless planar beach, the shear of any generated long-shore current will be weak so that shear wave energy may be undetectable. This explains why shear waves were not discovered until field experiments were carried out on barred beaches.

The instability by which shear waves are generated has a number of other analogs in nature. Large-scale coastal currents, flowing along a continental shelf, have been shown to develop similar instabilities although with very larges scales. Similarly, under wave motions, the bottom boundary layer, matching the moving wave oscillations of the water column interior with zero velocity at the fixed boundary, is also unstable.

It can be shown that a necessary condition for such an instability is the presence of an inflection point in the velocity profile (the spatial curvature of the current field changes sign), a requirement satisfied by long shore currents on a beach. In that case, cross-shore perturbations will extract energy from the mean long-shore current at a rate that depends on the strength of the current shear. Thus, these perturbations will grow to become first wave-like meanders, then if friction is not strong, to become a field of turbulent eddies. The extent of this evolution (wave-like versus eddy-like) is not yet known for natural beach environments.

Shear waves can clearly form an important component of the near-shore current field.) Root mean square (RMS) velocity fluctuations can reach $35\,\mathrm{cm\,s^{-1}}$ in both cross-shore and long-shore components of flow. This corresponds to an RMS swing of the current of $70\,\mathrm{cm\,s^{-1}}$, with many oscillations much larger.

[2]Swash oscillations are commonly expressed in terms of their vertical component.

Conclusions

As ocean waves propagate into the shoaling waters of the nearshore, they undergo a wide range of changes. Most people are familiar with the refraction, shoaling and eventual breaking of waves in a near-shore surf zone. However, this same energy can drive strong secondary flows. Wave breaking pushes water shoreward, yielding a super-elevation at the shoreline that can accentuate flooding and erosion. Waves arriving at an angle to the beach will drive strong currents along the beach that can transport large amounts of sediment. Often these currents form circulation cells, with strong rip currents spaced along the beach.

Natural waves occur in groups, with heights that vary. The breaking of these fluctuating groups drives waves and currents at the same modulation time-scale, called infragravity waves. These can be trapped in the nearshore by refraction as edge waves. Even long shore currents can develop instabilities called shear waves that drive meter-per-second fluctuations in the current strength with timescales of several minutes.

The apparent physics that dominates different beaches around the world often appears to vary. For example, on low-sloping energetic beaches, infra-gravity energy often dominates the surf zone, whereas shear waves can be very important on barred beaches. In fact, the physics is unchanging in these environments, with only the observable manifestations of that physics changing. The unification of these diverse observations through parameters such as the Iribarren number is an important goal for future research.

See also

Breaking Waves and Near-surface Turbulence. Coastal Circulation Models. Coastal Trapped Waves. Beaches, Physical Processes Affecting. Sea Level Change. Surface, Gravity and Capillary Waves. Wave Generation by Wind.

Symbols used

E	wave energy density
H	wave height
H_s	significant wave height
L	wavelength
L_0	deep water wave length
P	wave power or energy flux
R_s	significant swash height
RMS	root mean square statistic
S	radiation stress (wave momentum flux)
T	wave period
$\bar{V}$	mean longshore current
X	distance coordinate in the direction of wave propagation
a	wave amplitude
a_s	wave amplitude at the shoreline
c	wave celerity, or phase velocity
c_g	wave group velocity
g	acceleration of gravity
h	water depth
k	wavenumber (inverse of wavelength)
n	ratio of group velocity to celerity
m	ratio of infragravity swash height to offshore wave height
n	edge wave mode number
u	water particle velocity under waves
v	long-shore component of wave particle velocity
x	cross-shore position coordinate
y	long-shore position coordinate
z	vertical coordinate
β	beach slope
γ	ratio of wave height to local depth for breaking waves
η	sea surface elevation
θ	angle of incidence of waves relative to normal
ξ_0	Iribarren number
ρ	density of water
σ	radial frequency ($2\pi/T$)
φ	cross-shore structure function for edge waves
$\bar{\eta}_{max}$	mean set-up at the shoreline
∇	gradient operator

WEDDELL SEA CIRCULATION

E. Fahrbach and A. Beckmann,
Alfred-Wegener-Institut für Polar- und Meeresforschung, Bremerhaven, Germany

doi:10.1006/rwos.2001.0371

Introduction

The Weddell Sea is an area of intense air–sea interaction and vertical exchange. The resulting cold water masses participate in the global thermohaline circulation as deep and bottom waters.