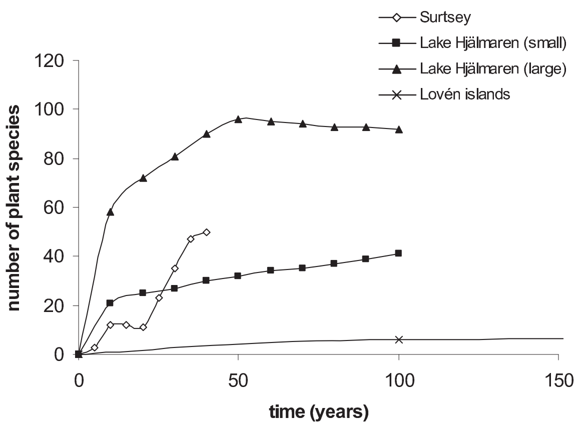

FIGURE 1 Number of species as a function of time since the island appeared, for Surtsey (Iceland), small and large islands in Lake Hjälmaren (average values) (Sweden), and the Lovén Islands (Norway).

accumulation and community assembly will be slower on more isolated and smaller islands, in comparison to large islands at short distances from the mainland.

General ecological conditions on an island at the start of the succession (e.g., climate and initial soil conditions) have been found to strongly determine the rate of succession. Increasing temperatures and precipitation enhance rates of organic matter decomposition and soil formation. Succession and species accumulation rates on boreal islands will hence be slower than on islands at tropical latitudes, because processes of weathering and soil formation are slower. Moreover, the potential species pool is smaller at higher latitudes. Additionally, characteristics of the substrate will determine the successional rate. Parent material with a fine texture will allow a faster succession because it is likely to weather more rapidly, increasing phosphate availability and the formation of an appropriate soil structure.

Fig. 1 illustrates the species accumulation rates of three island examples in the boreonemoral and boreal climate zone of the northern hemisphere. The 40 islands of Lake Hjälmaren (482 km^2, south-central Sweden, 60° N) appeared after a lowering of the water table in 1886, and ranges in size between 20 m^2 and 1 km^2. The islands can be divided into large (greater than 0.3-ha) and small (less than 0.3-ha) islands. As can be seen, large islands reach equilibrium at a time scale of about 50 years, after which species richness slightly decreases, whereas small islands still continue to accumulate species, and the equilibrium may last for more than a century. The number of species at small islands is at every point in time always smaller compared to that of large islands. The Lovén islands (78° N), situated in Spitsbergen, Norway, were successively released from beneath the ice during the regression of a valley glacier. The islands are about 3 km from the mainland and separated by at least 1 km of open water. Because these islands are situated in the boreal climate zone, species accumulation is very slow, and only six species were established on the island after a period of 100 years. Surtsey (63° N) covers 1.4 km^2 and is situated at the southern coast of Iceland; it appeared after a volcanic eruption in 1967. The slower species accumulation rate on the Lovén islands and Surtsey, in comparison with the islands in Lake Hjälmaren, despite the similar latitudinal position, may be attributed to the much higher degree of isolation of this oceanic island in comparison with the lake islands. Plant species richness increased for the first 15 years and then stabilized. The steep increase of species richness after 20 years is due to the establishment of a seagull colony on the islands. Seeds are transported by the birds, and there is a spatially concentrated input of nutrients, allowing a larger subset of species to establish. This clearly illustrates the importance of chance events in early successional stages on islands.

SEE ALSO THE FOLLOWING ARTICLES

Continental Islands / Convergence / Dispersal / Species–Area Relationship / Surtsey / Vegetation

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SURF IN THE TROPICS

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The wave climate around islands is highly variable in space and time as a result of the level of exposure to incident waves and the frequency of local and distant storm events. The magnitude of the incident waves affects the distribution of benthic communities such as algae, corals, and fish assemblages, as well as the distribution of sediment and the accretion and erosion of shorelines. Topographic effects often provide waves highly sought by surfers. Waves breaking on coral reefs around islands can have a significant impact on sea level and mean currents in an otherwise sheltered lagoon, flushing water, nutrients, and pollutants through the system and helping to maintain a healthy marine environment.

LINEAR WAVE THEORY

Field research carried out on tropical islands over the past several decades has led to fairly robust understanding of wave transformation, water levels, and circulation on fringing reef systems. The basics of the wave physics are the same as for open coast sandy beaches found worldwide; however, the importance of bathymetric irregularities and characteristic geometries of fringing reef systems on wave and current dynamics renders the circulation significantly different. The close link between the circulation and the health of the living reef ecosystem is recognized. Linking sediment and solute transport to predictive models of wave transformation and wave-driven flow with wind and tidal forcing is a difficult problem, but much progress has been made. The following sections discuss basic elements of wave physics and the dynamic consequences of waves impinging on a typical fringing reef system.

Wave Characteristics

Linear wave theory adequately describes many of the observed features of wave propagation and transformation, and the reader is referred to the list of references following this article. A schematic of an idealized wave is shown in Fig. 1, where wave height is the distance between wave crest and trough, wave period is the time taken for a wave crest at position B to reach position A, and wavelength is the distance between consecutive wave crests. Wave periods are typically in the range of 5 to 20 seconds, and the corresponding range of wavelengths in 20 m depth is 39 m to 270 m.

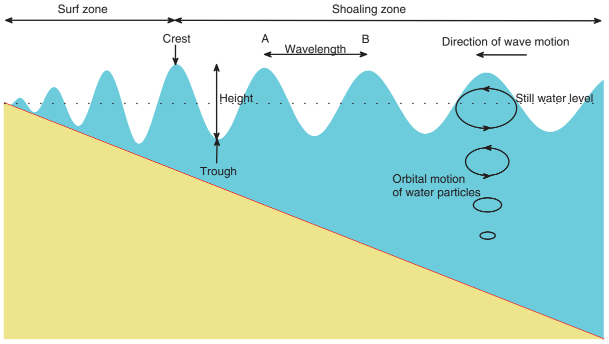

FIGURE 1 Schematic of the transformation of ocean surface waves approaching a shoreline. Through the shoaling zone, waves slow down, their wavelength decreases, and their height increases as they approach the surf zone. Through the surf zone, wave height is attenuated by dissipation processes and decreases to near zero at the shoreline.

Waves in the open ocean are generated by winds. In general, wave growth depends on the strength of the wind, the duration over which the wind blows, and the region (fetch) that the storm covers. As waves propagate outside their generation region, they begin to disperse, with longer-period waves traveling faster than shorter-period waves. As the waves sort themselves out, they form wave packets, or groups. In a typical group, there are anywhere from 5 to 10 wave crests that vary in height from small amplitudes at the front and rear of the group to larger amplitudes in the middle. The group pattern travels at one-half the speed of the individual waves in deep water; thus, the individual waves pass through the groups as the group pattern propagates shoreward. The energy of the ensemble of all waves in the wave field propagates with the group. In the absence of wave breaking or bottom drag, the transmission of wave energy, called the energy flux, is conserved for a given length of wave crest.

It is important to recognize that waves represent a propagation of energy, not mass, and that the energy propagates with the wave group rather than the individual waves. Beneath the surface, individual water particles describe almost closed ellipses (or circles in deep water) as shown in Fig. 1. As the wave crest passes, the particles at the top of the elliptical orbit are moving in the direction of wave motion, and as the trough passes, the particles at the bottom of the elliptical orbit are moving opposite to the direction of wave motion. Fig. 1 also illustrates a number of changes that occur as waves enter shallow water. Through the shoaling zone, the speed of the wave crest decreases with depth, and because the wave period must remain the same (to conserve the number of waves in a given time span), the wavelength must also decrease. In order to conserve energy flux, there must be a corresponding increase in wave height, thereby increasing its steepness. Eventually the wave crest overturns, or breaks, onto the front face, creating turbulent vortices and dissipating energy, and thus the wave height diminishes through the surf zone, approaching zero at the shoreline. Waves with residual height are reflected off the beach face and head back out to sea.

Wave Refraction

The dependence of the speed of the wave crests on the water depth leads to the process known as refraction.

Consider a very long crested wave approaching an island. That part of the wave crest that first encounters the shallow water around the island will slow down, whereas the part of the crest out in deeper water will be moving faster and will thus move ahead of the slower-moving part of the wave. This process is called refraction, and it causes the wave crests to bend toward the island, wrapping into what would otherwise be a shadow region behind the island. However, as the waves refract, they are effectively being stretched along the crest, and conservation of energy causes the wave height to decrease such that the wave energy flux (proportional to wave height squared times the group velocity) per unit length of wave crest remains constant. The greater the angle between the direction of the refracted wave and the corresponding deep water direction, the greater the reduction in wave height. State-of-the-art numerical models do a reasonably good job of simulating wave refraction over complex bathymetry, although care must be taken to properly represent intersecting waves refracting around opposite sides of an island.

Wave Momentum

Consider a column of water with a horizontal area of 1 m^2 extending from the sea surface to the bottom. Under the wave crest, this water column has a greater mass than it has under a trough. Momentum, given by mass times velocity, is thus greater under the crest than under the trough, and thus, when averaged over a wavelength, results in a flux of momentum in the direction of wave motion. This momentum flux caused by the presence of surface waves is known as radiation stress. As waves approach a shoreline or a sufficiently shallow reef, they eventually break, dissipating energy and transferring momentum to the water column. To conserve momentum, the gradient in wave momentum translates to a force that can drive a mean current in the direction of the waves or that is balanced by an equal and opposite force, such as a pressure gradient resulting from water piling up against a shoreline. On coral reefs around tropical islands, these wave-driven flows can be quite strong. A schematic representation of wave-driven flow over a submerged reef is shown in Fig. 2. Through the surf zone, the gradient in wave momentum drives a current in the direction of wave motion. However, when the waves reach the top of the reef slope, they cease to break, the gradient in wave momentum vanishes and is thus unable to provide a force to drive the flow over the reef top. In this one-dimensional case, assuming the current is uniform throughout the water column, conservation of mass requires the current to increase as

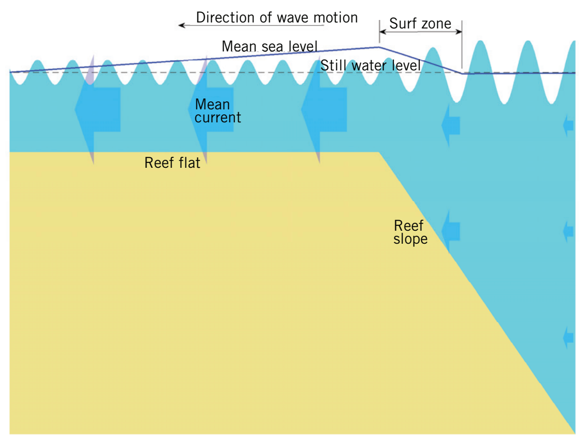

FIGURE 2 Schematic of wave-driven mean flow (large arrows) and corresponding mean sea level (solid blue line) over a submerged reef. Through the surf zone, the mean current is forced by radiation stress gradients and an opposing pressure gradient associated with the sea surface slope (blue line). Seaward and shoreward of the surf zone, the current is forced only by the pressure gradient caused by the sea surface slope.

the depth decreases, as illustrated in Fig. 2 by the larger arrows representing stronger currents. A pressure gradient is established by increasing mean sea level through the surf zone to a maximum at the top of the reef slope. From there, the mean sea surface slopes down to the level of water in the lagoon, providing a pressure gradient sufficient to drive the flow over the reef crest. Through the surf zone, the slope of the mean sea surface opposes the cross-reef flow such that increasing the mean sea surface elevation at the top of the reef slope increases the flow across the reef flat while decreasing the flow through the surf zone until the mass transport over the reef flat is equal to the transport through the surf zone. The magnitude of the cross-reef flow depends on the magnitude of the incident waves offshore, on the width of the reef flat, and on the depth of water over the reef flat. In the idealized case shown in Fig. 2, the mean sea level decreases to the still water level at the downstream, or lagoon, side of the reef flat. Observations have shown that in the presence of waves breaking on a reef, the mean sea level in the lagoon may also increase above the still water level, effectively reducing the cross-reef pressure gradient and cross-reef flow. It is possible for this wave-induced setup to raise the lagoon sea level sufficiently to prevent any cross-reef flow. Because of the typically narrow continental shelf around islands, wave setup can cause significantly larger sea-level fluctuations than can wind-induced storm surges that occur along continental margins.

As waves wrap around an island, refraction does not always orient the wave crests entirely parallel with the bathymetry, particularly on islands where the shelf is very steep. In this case, obliquely incident waves will also drive mean flows along the reef parallel to the depth contours.

WAVES ON REEFS

Natural coral reefs are significantly more complex than the idealized case shown in Fig. 2, with highly variable bathymetry across and along the reef and occasional deep passages cutting through the reef. Lord Howe Island (31°34′ S, 159°05′ E), shown in Fig. 3, is situated 630 km off the east coast of Australia, and is approximately 10 km long and up to 2.6 km wide. It is roughly crescent shaped, with a lagoon facing southwest and bounded by a fringing reef about 6 km long; it is the southernmost coral reef in the world. The lagoon and reef are shown in Fig. 3, with depth contours in gray; the white regions associated with wave breaking indicate the location of the reef crest. The reef is cut by three deep passages connecting the lagoon and open ocean, one at the northern end of the reef (North Passage) and two at the southern end (Erscotts Passage and South Passage). Mean currents, obtained by averaging 2-Hz data over five minutes, were recorded continuously for a three-week period at a number of sites indicated on Fig. 3. The current data shown in Fig. 3 have been smoothed and decimated to 12 hourly vectors, removing most of the semi-diurnal tidal component. These data confirm the presence of a mean inflow over the reef crest as predicted by the idealized one-dimensional model described above. However, in the two-dimensional case, once the flow enters the lagoon, it may be deflected alongshore toward the deeper passages, where the absence of wave breaking allows the water to exit back out to sea. The current pattern shown in Fig. 3 indicates a persistent outflow in the passages. Current speeds in the lagoon are generally weaker, with occasional short pulses of strong currents. On a natural reef, current magnitude is also limited by bottom friction governed by the roughness of the bottom. Observations from Kaneohe Bay, Oahu, Hawaii, have shown that the dissipation of wave energy by bottom friction is similar to depth-induced breaking over a very wide reef flat.

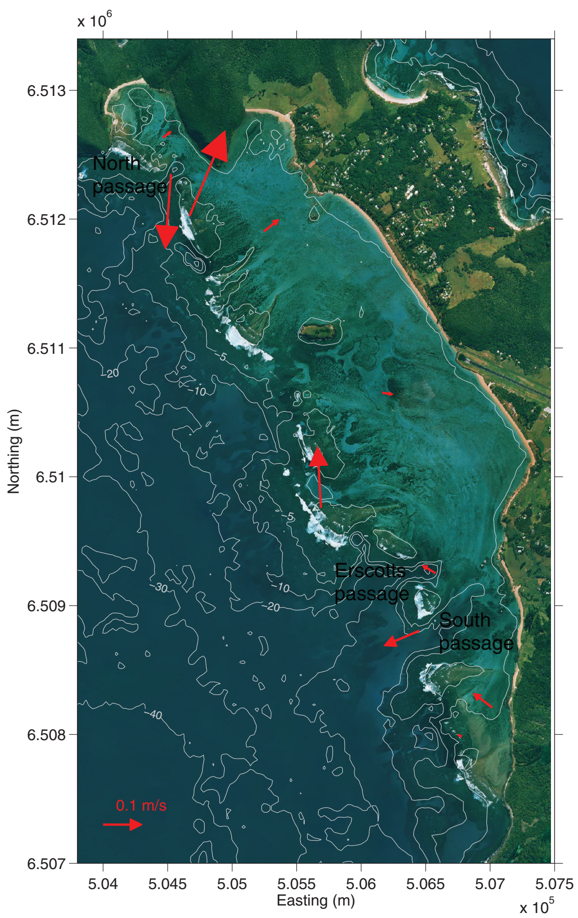

FIGURE 3 Mean current vectors measured along the reef and in the lagoon at Lord Howe Island, November 27 to December 14, 2004.

Clearly, there is variability in current speed and direction not always consistent with wave forcing, and it may reflect contributions that are the result of other processes such as wind and tidal forcing, which will certainly dominate during periods of low waves. Application of a state-of-the-art numerical model reveals the complex spatial variability of the wave-driven currents, as shown in Fig. 4. In this case, the model is forced by a constant wave height and direction at the offshore boundary with no tidal or wind forcing; thus, the spatial variability in the modeled flow field is mainly determined by spatial variability in the bathymetry. Although the model shows inflow over the reef and outflow in the major channels, qualitatively similar to the observations shown in Fig. 3, bathymetric low points along the reef crest are also sufficient to cause outflow.

The magnitude of the cross-reef flows and the relative shallowness of many coral reef lagoons leads to effective flushing of the lagoon with time scales on the order of the semi-diurnal tidal period or less. Tropical waters surrounding coral reefs are generally nutrient-poor, but the rapid flushing maintains a high through-flow of low-nutrient water sufficient to maintain the highly diverse coral reef ecosystem. Nutrient uptake by some corals and benthic communities has also been found to increase with increasing flow rate. Wave-driven flows also wash away fine suspended sediment from terrestrial runoff, which can have a detrimental effect on corals by reducing

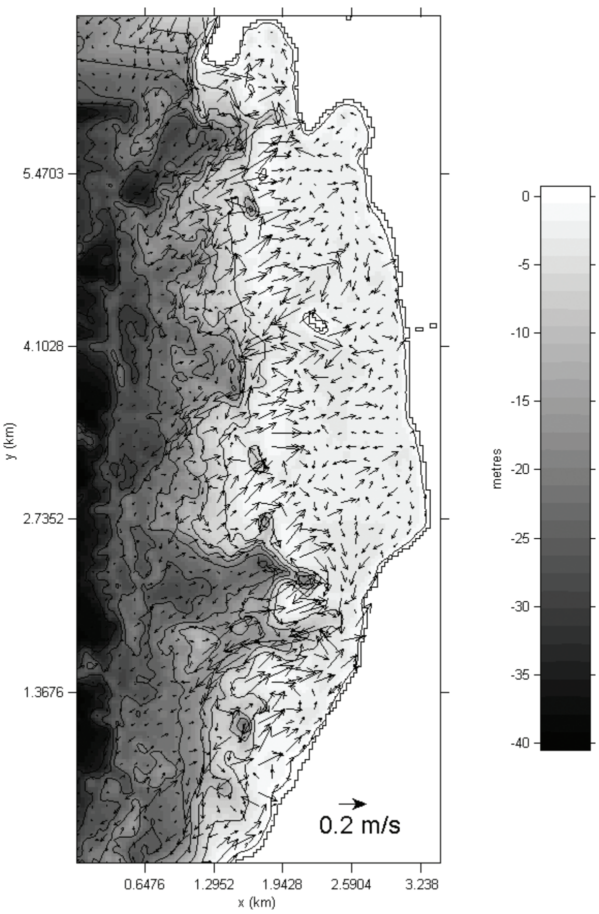

FIGURE 4 Numerical simulation of wave-driven currents at Lord Howe Island using a depth-averaged two-dimensional hydrodynamic numerical model (2DBeach, http://www.asrltd.co.nz). Background shading from light to dark indicates increasing depth according to the scale on the right.

ambient light levels and smothering benthic communities. Terrestrial runoff can also contain pollutants such as fertilizers, and rapid flushing of these pollutants helps prevent toxic algal blooms from developing in the lagoon. Most fish species have a larval stage, during which time dispersal and subsequent settlement are largely governed by local and large-scale current patterns.

SEE ALSO THE FOLLOWING ARTICLES

Beaches / Lord Howe Island / Reef Ecology and Conservation / Sea-Level Change / Tides / Tsunamis

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SURTSEY

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Off the southern shore of Iceland in the North Atlantic Ocean is a group of 14 islands and a number of skerries called Vestmannaeyjar or the Westman Islands. The largest of these is Heimaey at 11.6 km^2, the only populated member in the archipelago. The youngest and the outermost island is Surtsey. It was formed during a volcanic eruption that started on November 14, 1963.

AN ISLAND CREATED

That morning a fishing boat was situated some 20 km southwest off Heimaey. At dawn the captain saw a column of smoke coming out of the sea and realized that a volcanic eruption had started.

The pillar of smoke rose ever higher with pulsating explosions, constantly ejecting cinders and ashes, while bigger lava bombs were being flung up and splashed into the sea. The ash particles blown into the atmosphere were positively charged, and shortly after an explosion from the volcano, they triggered great lightning displays that were spectacular to watch and could be seen for miles around.

It took only 24 hours of volcanic activity for an island to take form, built up from the ocean floor of 130 m in depth (Fig. 1). This island was later named Surtsey after the fire giant from Nordic mythology, Surtur the Black—a name that suited the island's fiery origins and its black basalt lava formations.

SURTSEY SECURED FOR SCIENCE

Quickly recognizing the scientific importance of the new island, the Icelandic authorities decided to protect