
The Physics and Hydrodynamic Setting of Marine Renewable Energy

2

David K. Woolf, Matthew C. Easton, Peter A. Bowyer
and Jason McIlvenny

Abstract

Increasing interest is apparent in marine energy resources, particularly tidal and wave. Some TeraWatts of energy propagate from the world's oceans to its marginal seas in the form of surface waves (≈ 2 TW) and tides (≈ 2.6 TW) where that energy is naturally dissipated. The seas and coastlines around the UK and its neighbours are notable for dissipating a significant fraction of the global energy of waves (≈ 50 MW km⁻¹ on the Atlantic coast) and especially tides (> 250 GW north of Brittany). Displacing a significant fraction of the natural dissipation by energy capture is a tempting and reasonable proposition, but it does raise technical and environmental issues. Sustainable exploitation of the energy needs to consider diverse effects on the environment, waves and tides having a role in maintaining the shelf sea, coastal, estuarine and shoreline environment through associated advection, stirring and other processes. Tides are particularly significant in controlling the stratification of shelf seas and their flow characteristics. Surface waves are more important in determining conditions nearshore and in the intertidal zone. Also, the exploitation of wave and tidal resources is only practical economically and technologically at a limited number of energetic and accessible sites, and societal and ecological considerations inevitably narrow the choice.

Keywords

Hydrodynamics · Marine physics · Marine renewable energy · Shelf seas · Tides · Waves

Introduction and the Global Energy Context

The modern human's thirst for energy together with concerns over the influence of fossil fuel burning on atmospheric carbon dioxide, the "greenhouse effect" and climate has led to a wide search for alternative forms of energy. At the turn of the century, total global power use was estimated at about 12 TW

(Royal Commission on Environmental Pollution 2000), with 310 GW in the UK alone, equating to ~ 2 kW per person globally or 5 kW per person in the UK (which is fairly typical for an industrialized nation). These values are typically a factor of seven greater than electricity consumption alone and cover a diversity of needs, including transport and heating (Mackay 2008). As populations and appetites grow, both total and *per capita* energy use have risen and are expected to continue to rise, with consequences for fossil fuel reserves and the climate. A sustainable future requires that energy must be provided ultimately from naturally renewing supplies without severe environmental damage, and sustaining 12 TW from renewable energy resources is challenging. The total radiation absorbed by the Earth from the sun is $\sim 10,000$ times greater than this value, but it is distributed over the entire surface of the Earth and the direct conversion of solar energy (for example by photovoltaics to electricity) is not the

D. K. Woolf (✉) · M. C. Easton · P. A. Bowyer · J. McIlvenny
Environmental Research Institute, Centre for Energy
and the Environment, North Highland College,
University of the Highlands and Islands, Ormlie Road,
Thurso KW14 7EE, Scotland, UK

D. K. Woolf
International Centre for Island Technology, Institute of Petroleum
Engineering, Heriot-Watt University, The Old Academy, Back Road,
Stromness, Orkney KW16 3AW, Scotland, UK
e-mail: d.k.woolf@hw.ac.uk

only option and may not be the most practical either. Other forms of energy are more concentrated than the sun locally and are promising, but individually their practical exploitation cannot supply 12 TW.

A progressive electrification of most energy use including transport and heating is foreseen with sustainable energy sources directed mainly to generating electricity. Most studies of future energy supply envisage a major or complete supply of energy by a broad portfolio of sustainable energy conversion methods. The Special Report on Renewable Energy Sources and Climate Change Mitigation, SRREN (IPCC 2011) identifies six classes of renewable energy source and capture technology, of which ocean energy is one. Ocean energy is the internationally accepted term for a broad range of energy sources including surface and internal waves, tides, other ocean currents, ocean thermal energy and salinity gradient. Offshore wind is included in wind energy, another of the six classes, so for the purposes of this chapter, marine renewable energy (MRE) is here defined to consist just of ocean surface waves (hereafter, waves) and tidal energy.

In what follows, we focus first on the physics of tides and waves and the gross magnitude of the MRE resource, then look at the environmental function of tides and waves before considering the factors that lead to favourable circumstances for energy capture and the magnitude of the practical MRE resource. No consideration is made here of the detailed interactions with individual species or with ecosystems because they are discussed in other chapters and more broadly in the published literature (e.g. Shields et al. 2009, 2011; Scott et al. 2010; Burrows 2012), but we do discuss the role of waves and tides in processes such as transport, morphodynamics, stratification and dispersion that have ecological consequences.

Physics and Energy

Tides

Tides have been the subject of long and extensive study although, paradoxically, there have been relatively few studies in the energetic tidal channels where currents are strongest. Here the basics of tides are reviewed with emphasis on understanding energy balance and flux and on the conditions leading to favourable circumstances for energy exploitation. A broader and more complete description of tides is given by Pugh (1987), and Simpson and Sharples (2012) explain the major role of tides in the physical and biological oceanography of shelf seas.

Tides are forced globally by gravitational forcing (primarily associated with the Earth–Moon–Sun system). For example, the generation associated with the Moon results from an imbalance in two forces acting on elements of the

Earth (including parcels of water): the gravitational attraction of the moon and the centrifugal force associated with rotation about the combined centre of gravity of Earth and Moon. The combined equilibrium effect of these forces is towards the moon on the side of the earth nearest the moon, and away from the moon on the other side. This pattern underpins the “equilibrium theory of tides”, where an ellipsoid of fluid would rotate such that the major axis is always aligned towards the Moon. This equilibrium theory ignores the rotation of the earth beneath the tidal forcing, so is limited in predictive power because any fluid (including that in the oceans) on a rotating Earth cannot respond fully to the astronomical forcing that can be represented as a twice daily rocking of an ocean basin. However, it does provide a useful explanation for the dominance of periodicities in tidal flow that coincide with certain periodicities in the Earth–Moon–Sun system. For example, it can be inferred that two “bulges” in the equilibrium tide described above will pass a given point on the Earth’s surface in the period between consecutive instances of that point facing the moon. Those instances will be separated by slightly more than one day as the Earth needs to rotate slightly more than once to “catch” the orbiting moon. The frequency at which that alignment arises is the principal lunar semi-diurnal frequency, designated as the “M2 tide”. Hence, the M2 tide is semi-diurnal with a period slightly exceeding half a day. Similarly, the same alignment between a face of the Earth and the sun occurs precisely twice every day, and this semi-diurnal tide, the principal solar tide, is designated as S2.

Tides are forced at a number of frequencies characteristic of the Earth–Moon–Sun system. At most locations and for most purposes, the M2 tide is greatest in importance followed by the S2 tide, but there are locations (near M2 and S2 amphidromic points) where other tides are of similar or greater importance, typically the major diurnal tides. The S2 and M2 tides will be in phase twice per lunar month (29.6 days), so there is about a 2-week cycle between stronger spring tides (when M2 and S2 are in phase) and weaker neap tides (when M2 and S2 are in anti-phase). Ideally, all calculations relating to tides would be over very long periods (and based on measurements over those periods; Pugh 1987), but being more pragmatic, the absolute minimum is to resolve accurately the M2 and S2 tides, and to base calculations on the two-week spring–neap cycle (or more precisely, a semi-lunar month). Typically, the amplitude of both elevation and current for S2 will be one-third of the M2 equivalents, so there will be variation of typically a factor of two in both tidal ranges and current speeds between springs and neaps. Kinetic energy flux across a plane perpendicular to the predominant flow direction is the theoretical resource for very limited extraction of tidal stream energy and is proportional to the cube of current speed. Consequently, kinetic energy flux can vary by a factor of eight between springs

and neaps. The statistics of kinetic energy flux have to be calculated across a minimum of a semi-lunar month and use at a minimum information on M2 and S2 currents (conveniently, that is the maximum information that can be inferred from tidal diamonds on a chart). For major exploitation of tides, the kinetic energy flux can be very misleading (we will discuss this further in the Effects section below), and more useful though simplistic measures are the total energy and power in the tides. Energy and power are each proportional to the square of the amplitude of the elevation or current speed and will typically vary by a factor of 4 across a semi-diurnal month.

Useful indications of the energy (and resource) associated in total and with specific tidal components can be extracted from various studies of global and regional tides. Only two examples are summarized here. It is known that the work done by the moon on waters in the oceans is at the principal lunar semi-diurnal frequency, M2, whereas the sun mostly releases energy at S2. Generally, most dissipation will be in the same tidal components, although some energy can be transferred to different components (e.g. shallow-water tidal components) through various interactions. The tides are primarily generated in the deep ocean, but dissipate mainly after they cross the shelf edge. A total of 3.7 TW of work is done on the tides (2.5 TW at M2), of which 2.6 TW (1.8 TW at M2) is dissipated in the marginal seas (Munk and Wunsch 1998). These global values suggest that some 70% of the energy is contained in the M2 component. Apart from the M2 tide, most of the energy is at other semi-diurnal frequencies and at diurnal frequencies (periodicities close to one day). There can be local variations in the partition of energy between tidal components. For example, Robinson (1979) reports on calculations for part of a section in the Celtic Sea (on a line between the south coast of Ireland and the north coast of Cornwall, near the southwest tip of England) based on suitable observations over a 709-h period. Those calculations reveal that M2 accounts for >80% of the flux and that M2 and S2 together account for >96%. Below, we describe how the tidal resource can be understood in terms of regional tidal energy fluxes, taking the highly studied seas surrounding the UK as an exemplar.

In the modern era, radar satellite-borne altimeters have provided a means to map globally the dissipation of tidal energy (Egbert and Ray 2001). A small number of shelf sea areas is responsible for most of the dissipation of global tidal energy (Green 2010; Simpson and Sharples 2012). Tides and tidal dissipation are relatively high in the Atlantic and the seas bordering Europe, especially around the British Isles. The geographic resolution possible using satellite data is limited, however, and greater geographic resolution of fluxes and dissipation requires *in situ* instrumentation (or a numerical model adequately validated by data). Estimates of tidal energy flux and dissipation have a history dating back to cal-

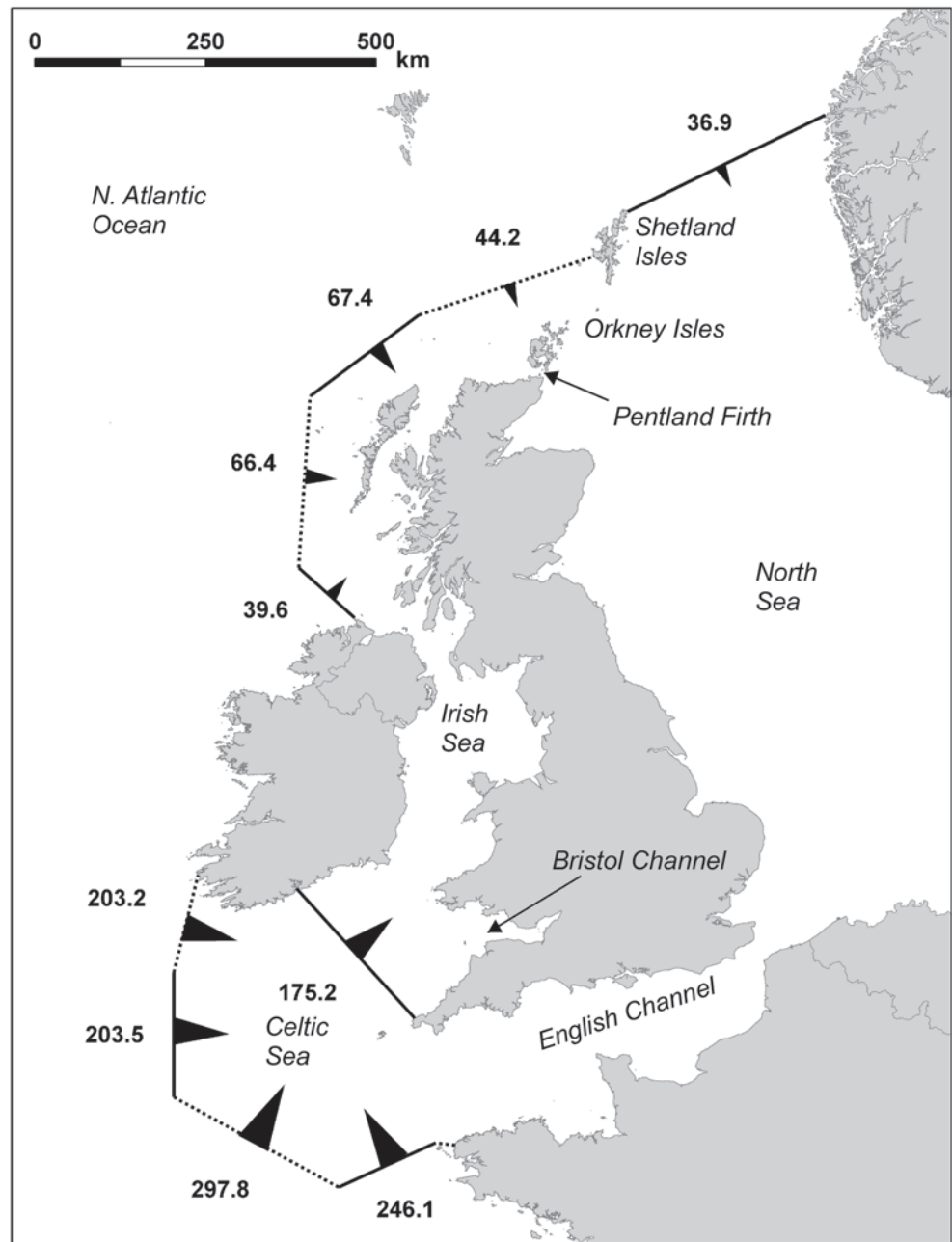
culations made for the Irish Sea by Taylor (1920), but they are limited to a few sea areas. The study of the seas around the British Isles is unusually thorough, and Cartwright et al. (1980) provide an excellent summary (see Fig. 2.1). They estimated M2 fluxes across a large number of sections surrounding the British Isles and additionally across several key sections within the bounded area, e.g. across Dover Strait. Many of the outer sections were defined close to the edge of the continental shelf, but necessarily some sections cross the shelf to convenient headlands. A southern boundary was defined by several sections from the tip of Brittany (Ouessant) to southwest Ireland (near Valentia). A flux of 190 GW is estimated across that boundary, with 45 GW penetrating to the northern part of the Celtic Sea, Bristol Channel and Irish Sea. Significant flux (16 GW) is estimated through the Straits of Dover to the southern North Sea, but most of the remainder (approaching 130 GW) has to be dissipated in the southern part of the Celtic Sea and English Channel. A northern boundary is defined between Malin Head (Ireland) and Florø (Norway), and an estimated flux of 60 GW across this boundary has to be dissipated in Scottish shelf and coastal waters and in the North Sea (note also the additional 16 GW into the North Sea through the Straits of Dover, which must also be dissipated).

The methods pioneered by Taylor (1920) and developed by Cartwright et al. (1980) and others greatly elucidate the flows of tidal energy, but they are relatively scarce. At a more local scale, an excellent example is given by Robinson (1979) who, within a study of the tidal dynamics of the Irish and Celtic seas, includes a fairly detailed study of energy flux vectors. More-detailed studies (following Robinson 1979) need to include minor energy interaction terms such as the work done locally by the currents on astronomical bodies, although the transports of tidal energy originating from the open ocean and the frictional dissipation terms are usually dominant. Such studies can be enlightening as one contemplates exploiting the energy.

As energy and power is the focus here, it is worth calculating the intensity of energy flows implied by the fluxes estimated by earlier workers. Hence, Fig. 2.1 depicts the average tidal power per unit length of each section (a convenient parameter to compare with estimated offshore wave-energy fluxes that are readily available and discussed below). At the southern boundary the fluxes are relatively large, between 200 and 300 kW m⁻¹. Fluxes across the northern boundary are weaker, though, consistently <70 kW m⁻¹, but even these lesser fluxes are generally similar to or greater than the wave fluxes at like locations.

A satisfactory comparison with the intensity of solar fluxes (typically several hundred W m⁻² across a horizontal plane during direct sunlight) is difficult to achieve, but on the face of it there is significant concentration of energy within tides, although the total global dissipation of tidal energy is

Fig. 2.1 Estimates of M2 tidal energy flux across the boundaries enclosing the British Isles, in kW m^{-1} (adaptation of a schematic developed by Cartright et al., 1980, who estimated values across key sections in GW, with values recalculated to power per unit boundary length to aid comparison with similar estimates for wave power; see Fig. 2.2)



tiny relative to solar energy. The global tidal energy dissipation (3.7 TW) is less than global energy use (12 TW in 2000, but rising thereafter), but considerable. Adding additional tidal harmonics to the 250 GW known for M2, the tidal energy flux from the deep ocean dissipated in the waters around the UK is coincidentally close to total energy use in the UK. Therefore, there is certainly sufficient tidal energy to be of interest, but given practical considerations, one should not expect it to solve the world's energy problems. We return to the "practical resource" later. Also, it can be inferred that the energy of the tides needs to be dissipated naturally within bounded areas, but tidal dissipation will be far from uniform and will be particularly great in some coastal areas, where

strong currents interact with the seabed. This issue is addressed further below after completing a more general summary of tides.

The tidal energy fluxes across the shelf edge west of Europe can be viewed as the limb of an Atlantic tidal system. As described above, those tides and the tides in all the world's oceans are a response to astronomical forces, but that response is modified from the equilibrium tide by inertia and the rotation of the Earth. Simpson and Sharples (2012) provide a cogent description of tides in shallow seas, so the following description is merely a brief summary. The tides are generally described as Kelvin waves, which are long or shallow-water waves (i.e. wavelength \gg water depth)

affected by the Earth's rotation. If rotation were ignored, then ocean tides might be regarded as long waves reflecting at continental boundaries to form patterns of standing waves, with nodal lines of low tidal range interspersed with anti-nodes of large tidal range (Simpson and Sharples 2012). Note that a different set of patterns is formed by each tidal harmonic, although harmonics close in frequency (and hence wavelength) tend to have similar patterns. The effect of rotation (and the resulting Coriolis force) can be thought of as replacing the nodal lines with amphidromic points, or simply "amphidromes", with tides rotating around them and zero tidal amplitude at the amphidrome. An amphidromic system can be described by two sets of lines, "co-tidal" and "co-range". Co-tidal lines radiate from each amphidrome and join points of identical phase in tidal elevation (e.g. high water will coincide at each point on a co-tidal line). The radiating lines are spaced such that they describe the full rotation of the tide around the amphidrome in a single tidal cycle. Lines of co-range connect points of equal tidal range, and the innermost of these lines are concentric to the amphidromes so describe the increasing tidal range outwards from the amphidrome. Farther from the amphidromes, maps of range are complicated by interactions of neighbouring amphidromes and interactions with the continental margins. The tides on the continental shelf west of Europe are primarily a function of an M2 amphidromic system centred far away in the western part of the North Atlantic. The anticlockwise rotation of the M2 tide around that amphidrome implies tides propagating northwards up the European continental shelf. Tides north of the UK and in the North Sea are also strongly influenced by a M2 amphidrome east of Iceland. Basin-scale amphidromic systems in the seas around the British Isles further complicate the tides around the UK.

The propagation of long-water waves depends on water depth. The phase and group velocity of such waves (and, for a given wave period, the wavelength) depend on the square root of the water depth. The tides, which are Kelvin waves with typical wavelengths of 8,000 km, are altered greatly where they cross the continental shelf edge from waters thousands of metres deep to shelf waters typically 100 m or so deep. Long waves propagate more slowly in shallow water and the motion is necessarily restricted by the water depth. As a result, tidal elevation (and tidal range) is typically elevated by a factor of 2.5 on the shelf, whereas tidal currents are typically elevated by a factor of 16 (Simpson and Sharples 2012).

Another significant effect of rotation apparent for tides (Kelvin waves) on the shelf and in coastal waters is an interaction with a coastline that amplifies the amplitude (and tidal range) near the coast. Consider a Kelvin wave propagating anticlockwise (the direction of rotation applicable to the northern hemisphere) around an amphidrome within an ocean basin. At the margins of that basin, the tide will propa-

gate with the continental margin on the right and the interaction will result in an amplification of tidal range adjacent to that margin (again, an effect attributable to rotation). The amplification will only fall off gradually with distance from the coast—the appropriate length scale is the Rossby radius (equal to the speed of the shallow wave divided by f , the Coriolis parameter), which is typically 250 km for the depth of shelf seas and >1,000 km for deep-ocean water—and the amplification is rather general for coastal and shelf water where the tide propagates with the coast on the right (i.e. in the northern hemisphere).

As noted above, Kelvin waves will reflect when they meet a coast at right angles to their direction of propagation, and the tides on the shelf will generally be a superposition of incident and reflected waves. At some sites (e.g. on the north coast of Scotland, where the tides propagate eastwards towards Pentland Firth and Orkney), the incident wave is dominant and the net effect resembles a simple progressive wave. In most cases, the reflected waves are highly significant and in some locations the superposition can resemble a standing wave with equal and opposed incident and reflected waves. An interesting case is the resonant system, where a natural periodicity dictated by the coastline and bathymetry coincidentally matches the periodicity of one of the major ocean tides. The resonant condition is that the distance between the shelf edge and the closed end of a gulf (or any bay or closed channel) should be an odd multiple of the quarter wavelength of the particular tide. In principle, the tidal range within a resonant system can grow indefinitely as it is fed by energy from the ocean tide, but in practice the tidal range and currents will be large but finite such that the energy input is balanced by frictional dissipation. The most famous example of this is in the Bay of Fundy, where the quarter-wavelength of the M2 tide almost precisely matches the length of the bay and tidal ranges up to 16 m are experienced. The Irish Sea is an example of a system that is not far from resonance and the tides are also very large, but which could in principle be nudged closer to resonance by civil engineering works (Pugh 1987).

Waves

Ocean wind waves and swell are a result of the action of the wind on the sea surface. A wind sea is the response to contemporary winds, and swell is the relic of earlier winds. The distinction between a wind sea and a swell is not always clear, but it is not essential because the same physical laws and formulae apply to both. Both types of waves arise at a range of wavelengths, but although a wind sea will include short ripples, swell is restricted to longer waves of tens or hundreds of metres wavelength. Waves propagate across the sea surface, but the motion of fluid elements within a wave

is primarily orbital, with the frequency of the wave and a surface diameter equal to the wave height. Water far below the sea surface will also follow an orbit but with a monotonically decreasing diameter with increasing depth. In shallow water the orbits are compressed in the vertical, first to an ellipse and finally to a line near the seafloor. Some of the energy of a wave is present and can be extracted from far below the sea surface, but a diminishing fraction.

The generation of the wind is a complex process resulting from solar heating and the Earth's rotation, and a fraction of the energy of the wind over the ocean surface is cascaded into wave energy. Very little of the Sun's power is diverted into waves (only exceeding 1 W m^{-2} in very strong winds). The potential utility of wind waves for sustainable energy arises first from the property of waves of accumulating energy over hours or days, and second, in the case of ocean swell, from energy being stored efficiently for several days (in the absence of strong wind forcing) and being propagated thousands of kilometres over many days to the ocean margins and an accessible location for exploitation. The physics of wave generation and propagation are largely neglected below, although we describe some formulae and results that help explain the practical value of wave energy. A specialist text on ocean waves (e.g. Holthuijsen 2007) provides a more-complete description of wave physics and statistics.

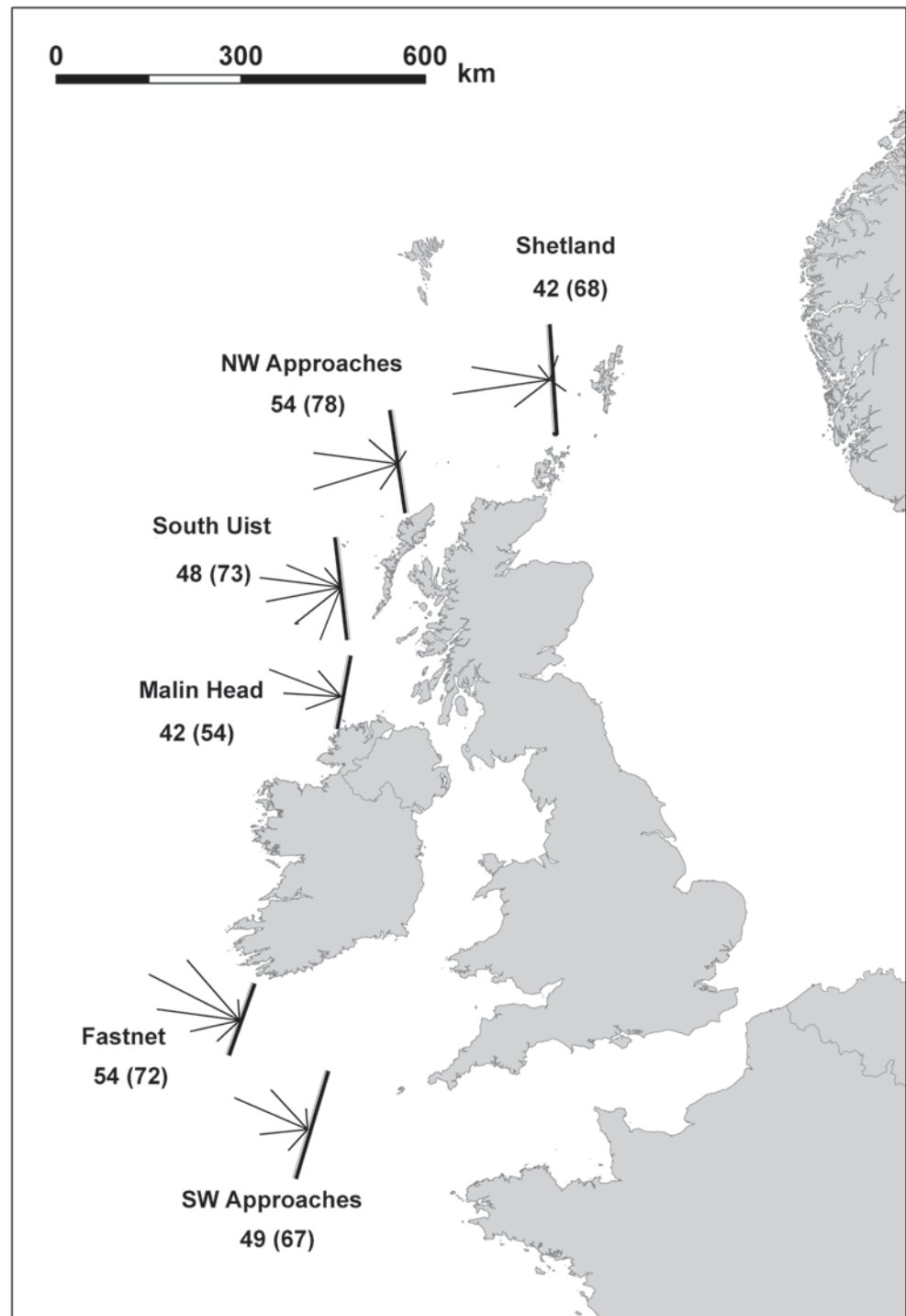
Where the wind blows steadily on the sea surface, waves may grow from nothing to a maximum, or fully developed wave height, over many hours, and the energy within the waves grows with the square of wave height. In addition to the increase in wave height, waves generally grow longer as non-linear interaction between waves of different wavelength directs most of the energy to longer waves. All waves move, i.e. they propagate, and longer waves move faster. A dispersion relationship relates the period and wavelength of a deep-water wave, with the wavelength proportional to the square of the period. A phase velocity can be defined as the speed of individual waves, and group velocity describes the speed of a packet of waves and of the transport of the energy within the wave field. The phase velocity is twice the group velocity for surface waves in deep water, and both are proportional to the period of the waves or to the square root of their wavelength. Therefore, in the open ocean, a wave field can be expected to grow and propagate downwind in response to a storm. If the wave field travels a similar path to the storm (ideally, if the group velocity of the dominant waves coincides with the propagation velocity of the storm centre), then the wave field accumulates energy from the storm winds. The flux of energy within a propagating wave front is given by the product of the energy density and the group velocity, so is proportional to the square of the wave height multiplied by the period. The wave field will lose energy by deep-water wave-breaking, or whitecapping, and at some point a terminal or fully-developed wave height will be reached, though only if the winds

are sufficiently sustained and there is enough sea room for the propagating wave group. Hence, both the duration of the winds and the fetch are important.

It is useful to complement the qualitative description above with some practical values, based here on sustained wind speeds of 10 m s^{-1} (fairly strong, but common) and 20 m s^{-1} (very strong winds, typical of an Atlantic storm). Some empirical formulae are used for the height and period of wind waves, as proposed by Carter (1982) on the basis of observations. The fully developed significant wave height (the average height of the highest one-third of waves) increases with the square of wind speed, reaching 2.5 m for 10 m s^{-1} winds and 10 m for 20 m s^{-1} winds, and the period of the dominant waves increases to $>7 \text{ s}$ and 14–15 s, respectively. These waves will travel with group velocities of $\sim 5 \text{ m s}^{-1}$ or $\sim 10 \text{ m s}^{-1}$, respectively, when fully developed and will require the wind speed to be sustained for $>20 \text{ h}$ or $>40 \text{ h}$ and over a fetch of 225 km or 900 km, respectively. The energy density of 2.5 m waves is $\sim 4 \text{ kJ m}^{-2}$ of sea surface and if that energy is accumulated in $>20 \text{ h}$, an average energy accumulation rate of only $\sim 0.05 \text{ W m}^{-2}$ is implied (note that this is a net rate and energy lost through wave-breaking is subtracted from the energy picked up from the wind, and that any spreading of the wave packet is also neglected). For 10 m waves, the energy density is $\sim 60 \text{ kJ m}^{-2}$ and the average accumulation rate is 0.4 W m^{-2} over 40 h. As noted already, waves accumulate energy slowly, but given persistent strong winds and enough sea room they eventually store a huge amount of energy that is rapidly propagated across the oceans.

The property of the wave field most directly applicable to sustainable energy is the flux of energy crossing a line parallel to the coast, or perpendicular to the main direction of propagation, in kW/m length of that line (see Fig. 2.2). From the values given above it can be calculated that a swell of 2.5 m height and 7 s period will contribute an energy flux of $\sim 20 \text{ kW m}^{-1}$, whereas a swell of 10 m and 14–15 s will contribute $\sim 600 \text{ kW m}^{-1}$. It is apparent from these values that even quite rare instances of very large waves may contribute considerably to the average energy flux, and that more-common waves of 1 or 2 m play a less important role in determining energy flux. That simple insight has a few implications. First, one would normally expect wave energy capture to be sited where a large ocean swell is relatively common (e.g. on the Atlantic margin of Europe where swell from Atlantic storms may propagate). Conversely, a coastal site in a windy area may be of little interest, if there are no or few instances where a wave field can propagate over several hundreds of kilometres to that site. Also, one should beware of simple estimates of average flux, because these may be affected by a few instances of very high flux (which may be entirely useless if they exceed the operating tolerances of the capture devices).

Fig. 2.2 Wave energy fluxes offshore of the UK (redrawn from Mollison 1991). The thick lines define arbitrary sections demarcating the UK offshore resource and the numbers denote the calculated wave flux across each section per unit length of that section in kW m^{-1} . The numbers in parenthesis neglect the angle of approach of the waves (effectively they are assumed to cross perpendicular to the section) and the more accurate values precede the uncorrected values. The thin lines schematically show the wave energy from each 30° sector contributing to the total



Many available wave-power statistics are presented as average flux. Note also that although we have presented some indicative values above based on simple unidirectional waves with a dominant frequency, real wave fields consist of a confusion of directions and wavelengths. In other words, a wave field should at a minimum be described by a directional wave spectrum that describes the distribution of energy across wavelengths and directions. A thorough treat-

ment of the theoretical potential of wave power should integrate across this wave spectrum in calculating the total wave power across a defined line. Calculation of practical resource should also consider thresholds for capture. For example, the calculation should account for saturation of capture devices at a design limit and should exclude instances where the wind speed or wave height is impractically high for safe operation. Historical estimates were often limited to approxi-

mate calculations that additionally relied on sufficient data. Some of the more-recent calculations are more complete, but are also based on limited data.

Wave-energy resources are typically quoted as the wave power reaching an ocean margin at or within the shelf edge (the offshore resource) or for technology suitable for shallow water or the shoreline, reaching a particular depth contour or the shoreline. As noted above, there is a natural dissipation of wave energy associated with whitecapping, but the useful resource is the energy contained in waves that reach the ocean margins. Some global estimates are quite vague. Gunn and Stock-Williams (2012) reviewed several historical estimates and completed their own relatively thorough calculation, estimating a global wave flux to a line 30 nautical miles offshore from a defined coastline of 2.11 ± 0.05 TW, consistent with earlier estimates on the order of 2 TW. Regional estimates of wave-energy resource, for example for the UK, for Ireland, or for the UK and Ireland together have been published over the past few decades. Early estimates calculate the gross power, which ignores the direction of propagation, whereas later estimates used directional spectra, and Mollison (1986, 1991), for example, used directional spectra from wave-model hindcasts (see Fig. 2.2). The gross power (i.e. neglecting the fact that many waves do not propagate perpendicular to the defined line) yields values of $50\text{--}80 \text{ kW m}^{-1}$ for the UK and most of the Atlantic seaboard of Europe, but 40 kW m^{-1} is typical accounting for direction. A wave-energy resource for the UK and Ireland combined of 72 ± 6 GW (Gunn and Stock-Williams 2012) is estimated. Mackay et al. (2010a, b) investigated the uncertainty in wave-energy assessment and identified potential errors attributable to both limitations of historical data and the variability of wave climate. It is clear that each estimate of resource can be biased by the method of calculation, by errors in the source data, or by sampling bias (i.e. the values will reflect the period when the data were collected).

Although a total of ~ 2 TW approaches global land-masses, this offshore resource reduces progressively as the waves shoal, so that the nearshore and shoreline resources are smaller. Favourable sites for wave energy tend to be where ocean swell approaches the coast. Once the swell in the open ocean has reduced significantly below the steepness of newly generated wind waves, it propagates with almost no wave-breaking or loss of power. The swell steepens on the continental shelf, because the group velocity of waves is less in shallower water, and may eventually steepen to a stage where shallow-water wave-breaking is a substantial sink of energy. Also, a turbulent boundary layer will be set up by the interaction of the seabed with the deep motion associated with the waves (an orbital motion that decays with depth), resulting in some frictional dissipation. Therefore, a slow loss of wave power may be noticed on the continental

shelf. Mollison (1986) reviewed a few estimates for water of intermediate depth (20–100 m) that suggest that a loss of 1 % per km may be typical.

As noted previously, wave power can vary dramatically. At a favourable site frequented by swell, there is likely to be $< 1 \text{ kW m}^{-1}$ of wave power at least 10% of the time. At the same time, peak wave powers of the order of 1 MW m^{-1} are likely, which will at best be superfluous and at worst catastrophically destructive. Wave energy will vary on the same time-scales as the weather (i.e. weather systems passing through every few days), but will also exhibit seasonal and interannual variations reflecting the variability of ocean winds. For example, the UK wave climate features higher, more-energetic waves in winter and shows strong variation between winters paralleling the variation between wet and windy winters and drier, calmer ones (Woolf et al. 2003). The interannual variation is associated with the North Atlantic Oscillation (NAO) and other large-scale modes of variation in regional climate (Woolf et al. 2002; Mackay et al. 2010b). The behaviour of these modes under climate warming and hence the statistical characteristics of Atlantic cyclones and other factors affecting the waves reaching the Atlantic seaboard of Europe (Wolf and Woolf 2006) is difficult to predict. The greater wave power in winter may be helpful because it coincides with greater energy demand in the UK. Unfortunately, however, interannual variations are unhelpful because wind, wave and hydro-energy generation will all be lower in relatively cold, dry winters. Clearly, therefore, future climate is a significant consideration in evaluating the commercial prospects of wave energy (Harrison and Wallace 2005), although changes in wave climate related to anthropogenic climate change are anyway likely to be dwarfed by natural variation within the lifetime of a wave farm (Mackay et al. 2010b).

Tides, Waves and the Environment

Shelf seas and coastal waters are an important environment that supports a rich ecosystem that is economically significant and enriches the human experience. Tides and waves are an important part of the physics of this environment and below we explain how the physics partly determines the environment and the nature of the ecosystems.

Tides

Some of the simplest effects of tides on both the water column and benthic habitats are related simply to tidal currents. It has already been said that tidal currents vary from one site to another. Fast tidal currents present problems both for benthic organisms that need to attach to the seabed or for organ-

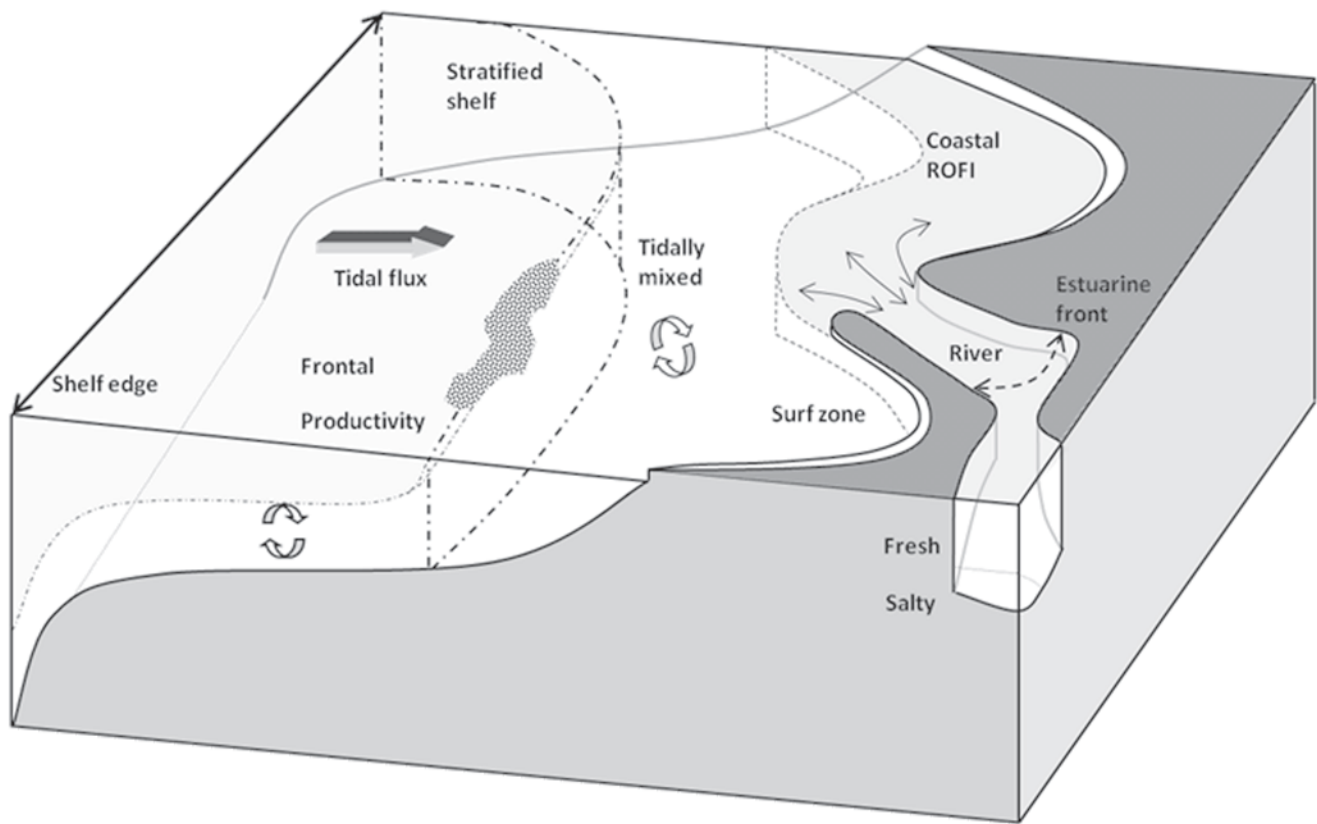


Fig. 2.3 A schematic of shelf and coastal seas, depicting a number of different environments. The effects of tides and waves on the environments are discussed in the text. ROFI refers to a region of freshwater influence, a region of coastal water where river water influences the stratification

isms within the water column that need to resist transport or dispersal. They may also be closely related to the seabed sediment, because in fast currents bare rock, or episodes of disturbance, removal and smothering, might be expected. Hence, speciation of fast-current regions should reflect adaptation to these specific conditions (Shields et al. 2011). Fast currents may also assist or hinder migration, with colonization depending on net transport over many tidal cycles (residual currents).

Apart from fairly direct responses to currents, most of the important effects are mediated through the influence of the tides (and to a lesser extent, the influence of wind and waves) on the zonation of the shelf seas into some distinct environments. The diversity of the physical environment from the shelf edge to estuaries is depicted in Fig. 2.3, and this illustration needs to be borne in mind when reading the paragraphs below.

Tides are an important part of the shelf-sea environment, are responsible for significant transport and most importantly stir the shelf seas. As illustrated in Fig. 2.3, there tend to be two distinct regions within the shelf seas, with fronts between them. Where the tidal current is strong and the water shallow, the sea will be mixed from top to bottom throughout the year. In other areas (deeper, or slower current regimes),

summer heating can partly isolate a warmer upper layer from a lower layer stirred by tides (seasonal stratification). Tidal currents certainly play a key role in determining the location of seasonal shelf sea fronts (Simpson and Hunter 1974; Simpson and Sharples 2012), and this can be understood first from the insight that vertical mixing depends on the stirring phenomenon. When the sea surface cools, the water column is overturned by convective instability, but when the surface warms, near-surface density reduces and the resulting buoyancy flux tends to stratify the water column, and that can only be overcome by vigorous stirring. The most effective stirring process is usually the stirring by turbulence induced by tidal flow over the seabed (we return briefly to the secondary roles of wind- and wave-induced stirring later). The power going into stirring by a tidal current cannot exceed the rate of dissipation of tidal energy and is proportional to the cube of the tidal current and a drag coefficient. For a typical drag coefficient, this dissipation rate will be $\sim 2.5 \text{ W m}^{-2}$ at a current speed of 1 m s^{-1} . For tidal stream sites, the dissipation rate is high, averaging over a tidal cycle up to the order of 100 W m^{-2} (Fig. 8 of Easton et al. 2012). These gross figures are misleading, however, because it appears that the power is not used efficiently in the mixing process. Simpson and Sharples (2012), for instance, estimate a mixing efficiency

of just 0.5% for one case, and it is assumed that most of the remainder generates heat in the turbulent boundary layer. Note also that the power (per unit depth) available to overturn the stratification will be inversely proportional to water depth, so where the geographic variation in buoyancy flux is small, it is the value of the ratio of water depth to the cube of the tidal current (most often averaged over a tidal cycle) that is critical in determining whether a water column is mixed throughout the year. As water depths are invariably shallower and tidal currents more often stronger inshore (also, in the coastal area, significant mixing is associated with features such as headlands and dissipation of wave energy), the most common pattern, depicted in Fig. 2.3, is for the water of the outer shelf to be stratified in summer, whereas closer to shore the water column is well mixed. The location of fronts between regions of mixed and stratified water can be effectively predicted from the ratio of water depth to tidal current cubed. In stratified water, the lowest layer will be effectively stirred by the tide, but upper waters will only be stirred intermittently by wind and the waves.

Coastal waters are generally well mixed, but this situation may be dramatically altered by river outflow, because freshwater can be an enormous source of “buoyancy flux”. Stratified zones exist within estuaries and coastal regions of freshwater influence, or ROFIs. Again, there is effectively competition between stratification by the buoyancy flux and stirring by tidal currents, and numerical models are quite effective in predicting the resulting zonation, given adequate bathymetry, tidal physics and estimates of river flow.

Various states of mixing and stratification affect phytoplankton and hence water-column ecology, because they affect the distribution of nutrients and the amount of light received. Frontal regions between fully mixed and stratified waters tend to be regions of convergence and vertical movement, so are conducive to high levels of primary productivity. Stirring may also influence encounter rates, feeding and reproduction (Shields et al. 2011).

Currents will also resuspend and transport sediment. In some areas dominated by fine sediments (e.g. within the Bristol Channel), resuspension of sediment can be so great as to make the water column opaque, lowering the light available for photosynthesis. Sediment accumulation will vary greatly according to sediment resuspension and transport, with consequences for benthic habitat.

A simple but important result of tides is a variation in sea level (tidal range), implying a variety of water depths and environmental conditions for benthic organisms below the low-water line and creating intertidal habitats. Both range and currents have a role in ecological zonation in benthic and intertidal habitats. For example, Burrows (2012) noted a statistical association between a shift from macroalgae to filter-feeders and tidal current speed, but only in areas where chlorophyll concentrations were high. As described below,

waves may be a more substantial cause of varying zonation nearshore and onshore than tides.

Waves

The influence of wind and waves on stirring the shelf seas is generally considered to be much weaker than the effect of tides. The relative strength of stirring can be understood from a calculation of the power dissipated per unit area of the sea surface associated with each process. Above, it was noted that tidal dissipation amounts to a power of about 2.5 W m^{-2} at a current speed of 1 m s^{-1} but can reach an order of 100 W m^{-2} at tidal stream energy sites. Simpson and Sharpley (2012) only discuss the direct effect of wind-stirring. They provide formulae for calculating the power going into turbulent motions from wind stress acting on a wind-driven current (of typically 2% of wind speed), and from these formulae, one can calculate a power of only $\sim 0.03 \text{ W m}^{-2}$ for a wind speed of 10 m s^{-1} . In other words, only hurricane-force winds are projected to have a similar stirring effect as tidal currents of the order of 1 m s^{-1} . There is evidence that the wind has a noticeable if secondary role in governing stratification (Simpson et al. 1978). The significant contribution of wind is partly attributable to a greater proportion of the energy from wind going into turbulent motion than is the case for tide (2.3 vs. 0.4%; Simpson et al. 1978).

Simpson and Sharpley (2012) did not consider the effect of waves on stirring and stratification, and this subject seems rather to have been neglected by physical oceanographers, although it does have the attention of coastal engineers (Nielsen 1992). Above, we noted that although the power going into waves is small, waves accumulate energy over large distances and durations. The accumulation and propagation of large amounts of energy (a flux of the order of 100 kW m^{-1} of wave front is common for ocean waves) implies that when and where this energy is lost to turbulent motion in the upper ocean, the strength of stirring must be considerable. Energy is lost from a wave field as ocean waves approach the coast through either direct frictional interaction with the seabed, or because the steepening of waves in shallower water results in wave-breaking (note that the longest ocean waves respond to the seabed even in depths of 100 m or more). Note again that Mollison (1986) suggested a loss of energy of 1% per km of wave may be typical in water of intermediate depth (20–100 m). Such a rate of loss implies a loss of 1 W m^{-2} from a wave field of 100 kW m^{-1} , suggesting that in wave-exposed shelf areas, the contribution of wave-induced stirring may be substantial. If an energetic wave field comes nearshore, then the rates of energy loss may be massive; for example, if a 100 kW m^{-1} wave field is wholly dissipated in a surf zone 100 m wide, then an average rate of dissipation of 1 kW m^{-2} is implied.

Marine Renewable Energy Technology and
Environmental Interactions

Shields, M.A.; Payne, A.I.L. (Eds.)

2014, IX, 176 p. 39 illus., 10 illus. in color., Hardcover

ISBN: 978-94-017-8001-8