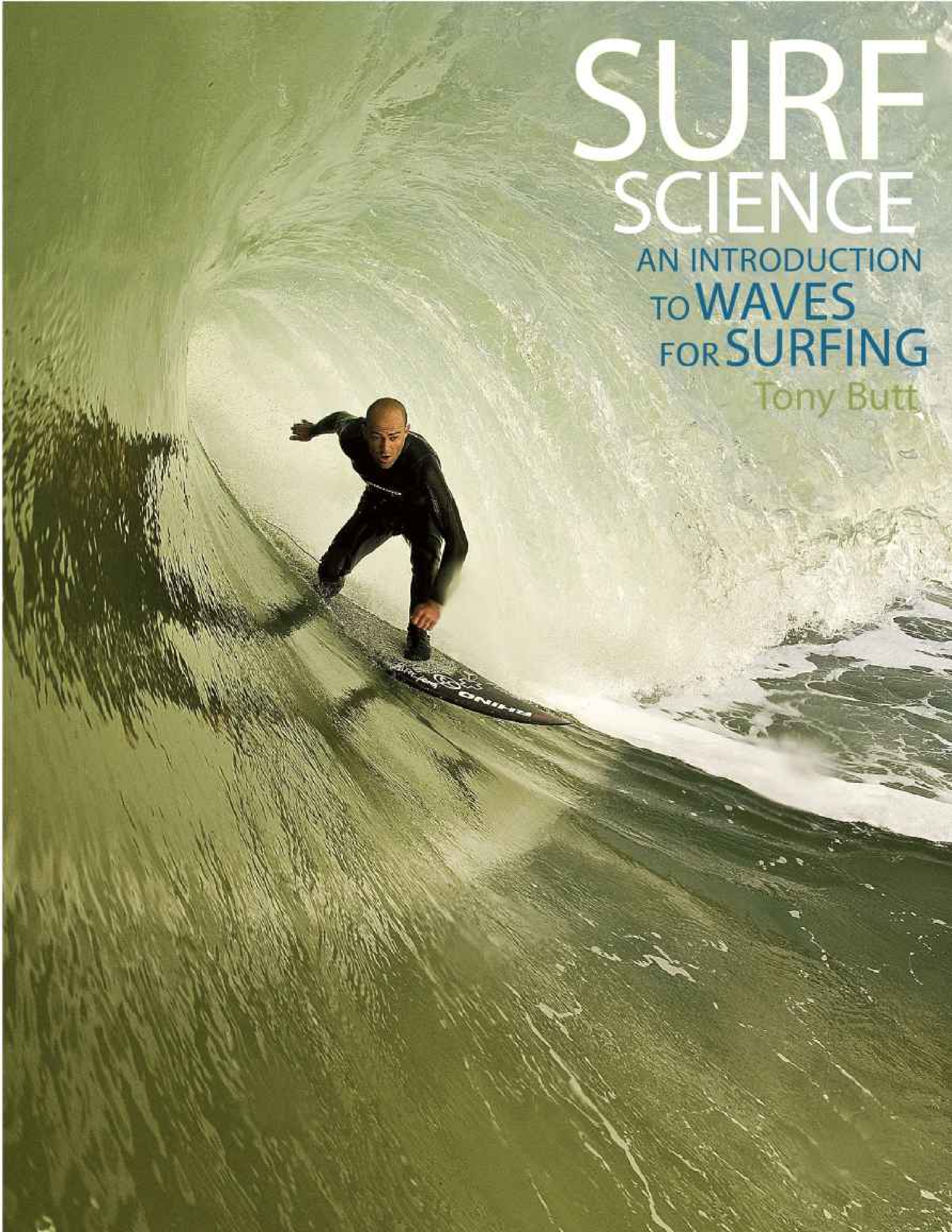
SURF SCIENCE

AN INTRODUCTION TO WAVES FOR SURFING

Tony Butt



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An Introduction to Waves for Surfing

TONY BUTT

Alison Hodge

About the Book

Have you ever wondered where surfing waves come from, what makes every wave different, why some peel perfectly and others just close out; why, some days, the waves come in sets and other days they don't, and how the tides, the wind and the shape of the sea floor affect the waves for surfing?

If you have, this book is for you. Now in its third edition, *Surf Science* is the first book to talk in depth about the science of waves from a surfer's point of view. It fills the gap between surfing books and waves textbooks, and will help you learn how to predict surf. You don't need a scientific background to read it – just curiosity and a fascination for waves.

What the critics said about earlier editions of *Surf Science*:

'one of the most sophisticated surfing books ever produced … a must for any surfer' (Ben Marcus, *Surfer's Journal*, May 2003)

'Tony Butt … is … a "surf forecasting guru". If he doesn't know it, it isn't worth knowing' (*Surf Europe*, Issue 22, 2003)

'So who would benefit from this book? Well, it's hard to think of a surfer who wouldn't.' (*The Surfer's Path*, Issue 33, 2002)

'an excellent resource for surfers who want a simple understanding of the how's and why's of wave creation and surf conditions' (hisurfadvisory.com, October 2004)

'jam-packed with many helpful graphs, diagrams, photographs' (*Longboard Magazine*, February 2005)

'clearly reaches out to both surfers and those interested in the surf… great color graphics and color photos' (David F. Narr, Associate Professor, College of Marine Science, University of South Florida, *Oceanography*, vol. 18, no. 2, June 2005)

About the Author

Tony Butt has a PhD in Physical Oceanography and worked with the Coastal Processes Research Group at Plymouth University for some ten years. He is a big-wave surfer and lives for most of the year in a forgotten corner of north-west Spain.

For my sister, Tina, 1947–2012

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Preface

If you've ever wondered where the waves come from, what makes every wave different and what factors affect the behaviour of a surf spot, you'll like this book. You don't need a scientific background to read it. In fact, you don't even need to be a surfer; all you need is a love for the ocean and a sense of curiosity.

It has now been ten years since I wrote the original version of this book. In this, the third edition, I have completely revised the text, revised the diagrams and photos, removed two chapters and added two new ones. So this edition is actually quite different from the last two.

The original idea behind *Surf Science* was to fill the gap between surfing books and textbooks. You'll find general surfing books with a basic oceanography section, easy to read but sometimes inaccurate; or advanced textbooks, much more complete but full of equations. However, you won't find much in between. With *Surf Science*, I have tried to steer somewhere between these two extremes, making the descriptions as complete as possible using familiar examples and light-hearted analogies, but without getting into any of the heavy maths. Of course, you can only go so far with that, so if you want to dig a bit deeper there are references to more advanced books.

The first seven chapters describe the life of a wave (or, more precisely, a packet of energy that briefly manifests itself as a wave), from before its birth in an oceanic storm to its final dissipation on the shore. Each of these chapters describes a different process and may be read on its own, but is probably better read as part of a complete sequence. Each of the next four chapters is a self-contained description of a phenomenon affecting the waves for surfing, and the last three chapters contain some useful knowledge on wave forecasting.

Most of the information in the original version of this book is still relevant, and so a lot of the changes I have made are just improvements in the descriptions. However, wave forecasting has changed quite a bit in the last ten years; so I have expanded and updated that section. You will also find the chapter on tides considerably longer than before, now doing a bit more than just scratching the surface of what is a very complicated subject. If you feel that the tides chapter is too advanced, you can leave it out without affecting the rest of the book. Finally, I have removed the chapters on wave climate and coastal morphology. These subjects are now covered in my book *The Surfer's Guide to Waves, Coasts and Climates*.

Tony Butt, Spain, 2014

1 Introduction

The paradox of impossible knowledge

'Looks like it's going southerly,' pointed out the surfer, as he gazed up at the clouds and sniffed the air.

'Yeah,' said the other, 'There's a warm front on the chart associated with a 985 sitting a few hundred miles west of Shannon.'

'The K2 buoy is already showing ten feet at fourteen seconds,' said the first. 'I reckon it'll pick up on the push of the tide – what is it anyway, springs or neaps?'

'It's a five point eight,' said the second, after consulting a strange list of numbers screwed up in his back pocket.

To the innocent bystander, this sounds more like scientific terminology than the jargon associated with some sport or leisure activity. Yet this is the sort of language surfers throughout the world – you and I – use every day of our lives. You see, without realizing it, most surfers are also scientists. Surfers who have spent many years riding different waves in different parts of the world have, without any special effort, become meteorologists, oceanographers, geographers, linguists and cultural experts. Through an obsession with tapping the ocean's energy to propel them along for a few seconds, surfers end up acquiring a large amount of peripheral information. All that watching, discussing, waiting and thinking gives us an insatiable thirst for knowledge, typical of scientists rather than sportspeople. This unique facet of surfing gives it a richness rarely found in other activities.

If you are a surfer, no doubt you will have asked yourself, while gazing out at the ocean or sitting in the line-up waiting for the next set, questions like these:

… Why is every wave different?
… Why are some waves more powerful than others?

… Why do some peel nicely and some just close out?
… Why, some days, do the waves come in sets of six, and other days in sets of three?
… How would this place work on a north, or indeed a south, swell?
… Why didn't that low produce any surf?
… Where did that swell come from?
… What are waves anyway?

And doubtless some of us have asked many more obscure questions. Some of these are readily answerable; others take some thinking about. Yet others – a surprisingly large number – are much more difficult or impossible to answer even at top oceanographic research level. There are some concepts that surfers know intimately, in a qualitative way, that are barely acknowledged by the scientific community, typically through lack of demand for practical application.

For example, the **groupiness** of a swell has been studied by some coastal engineers for the design of coastal structures. But exhaustive details of the length of time between sets, the number of waves in a set, whether each set has the same number of waves in it, or how the wave heights are distributed throughout the set, are not required for such a study. For surfing, however, these details can be crucial, especially on a big day.

Likewise, the answers to some questions that surfers are constantly wondering about, have been sitting there in the scientific literature for years. A little knowledge about wave periods, for example, can explain the difference between a strong, powerful new swell and a weak, gutless old one.

Apart from simply wondering why the waves behave the way they do, we probably spend a lot more of our time wondering what the surf will be like in the near future. This has a direct practical application. Since we plan our lives around the surf, being able to predict what the waves will be like tomorrow or next weekend might save us a lot of time and effort. Nowadays, surfers already have access to many efficient ways of predicting the surf. Using these facilities is becoming easier and easier, but if we don't have a rudimentary knowledge of meteorological and oceanographic terminology, and don't know the meaning of all those lines and symbols on the charts, we still run the risk of missing something. Sometimes, by just using those easy-to-read, highly-simplified forecasts, we are left wondering why the surf didn't quite turn out the way they said it would.

Many surfers are more competitive than they would like to admit. If you can sneak off and score perfect waves alone or with two of your buddies, while the rest of the surfing world is running around not knowing which beach to head for, then I'm sure you will. To stay ahead of the game, you are obliged – almost condemned – to know where to get the latest and best prediction, especially in places where crowding is an issue.

Within the world of wave and weather predictions lies a kind of paradox. If we knew exactly what it was going to be like every day, life might be a lot less interesting. Perhaps we actually *need* to be slightly in the dark to make things exciting. We need that degree of anticipation, that uncertainty, so that when that 'perfect' day does arrive, we appreciate it more.

Safe in the knowledge that we will never be able to predict the surf with one hundred per cent accuracy, we continue trying. In the process, we learn more and more about how waves work, what affects their quality and quantity, and what affects our ability to enjoy them; which, in turn, enhances our enjoyment of the waves and the natural environment in which they exist.

What's in this book?

If we want to know what a wave is and how it works, we need to know a little about its history. So the first seven chapters of this book describe a journey. But the traveller in this journey is not an individual wave; it is a 'packet' of energy.

The journey starts with the energy from the Sun entering our atmosphere. This energy drives our weather and wind, and generates waves on the ocean. It directs these waves towards our coasts while moulding them into different shapes and forms, and finally expends them on the coastline. The energy jumps from one form to another and, for a brief moment, is manifest in a sloping lump of moving water that we can ride on a board.

Once the Sun's energy enters the atmosphere, the atmosphere is set in motion by the uneven heating of the poles and the Equator, and that strange-but-vital ingredient, the **Coriolis force**. Large, swirling vortices of air, called low-pressure systems, form in the lower atmosphere over the ocean. The air circulating around these systems moves along the surface of the water, transferring energy from the air to the water and generating the first recognizable form of surfing waves: the **windsea**.

The energy is now contained in the surface of the ocean in the form of waves, which gradually begin to leave the generating area. The energy travels within the water in the form of **swell**, while the water itself moves in isolated, circular motions. The swell crawls across the ocean surface, gradually organizing itself into neat groups of undulations heading towards some coastline.

Once near the coast, these undulations are influenced by the topography of the sea-floor – the **bathymetry** – which bends and warps them into different shapes and sizes. When the waves get really near the coast and start to travel over shallow water, they break. It is here that the energy carried by the waves is transformed into moving water. And it is around here that we surfers can bleed off an infinitesimal part of this energy and use it to push ourselves along on our boards (Figure 1.1).

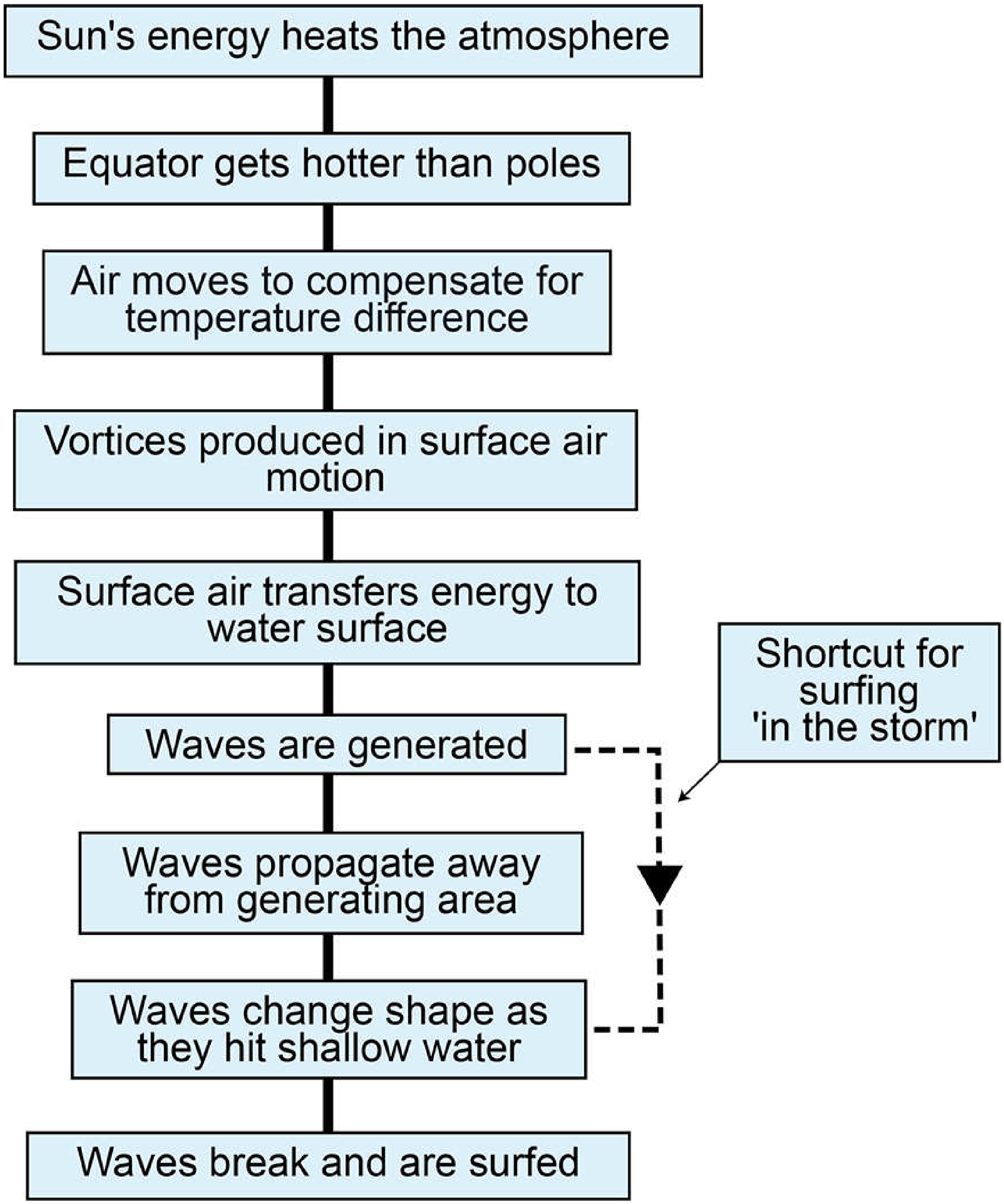

Figure 1.1: Flowchart summarizing the different stages in the journey.

The next four chapters of this book are a collection of self-contained topics, each describing an important coastal phenomenon that affects our surfing. A little knowledge of each of these subjects, in addition to some knowledge of the waves themselves, should enhance your surfing experience even more.

In some parts of the world, thanks to ocean currents and the arrival at the surface of cold water from below, the coastal sea temperature seems to be totally unrelated to the local climate. Chapter 8 explains why this is so, and discusses the temperature variation of surfing waters around the globe.

Sometimes, local winds on the coast generate waves that are not useful to us. These may ruin a nice clean swell by adding unwanted short, choppy windsea. The **sea breeze** – a local onshore wind that blows at certain times of the day in warm climates – can absolutely devastate perfect surf. Chapter 9 explains a few things about the sea breeze: how it works, and when and where it is most likely to occur.

Another factor that can mean the difference between waves and no waves at many spots around the world is the tides. Without a set of tide tables, many of us would be lost and confused. Thankfully, the tides can be predicted fairly accurately. Chapter 10 delves into some of the classical theories used to explain where the tides come from and how their movements can be predicted.

Turning up at some beach and not knowing where the rips are and in what direction they are running is just as bad as not having a set of tide tables. Chapter 11 looks at how rips work, and runs through a few common types of rips found at different types of surf breaks.

The last three chapters of the book deal with the important subject of wave forecasting. Without at least some knowledge of wave forecasting you could end up at the wrong beach while your friends enjoy epic surf somewhere else; or you could take the wrong day off work, and then spend the next day gazing at perfect surf out of your office window.

Chapter 12 looks at the basics of wave forecasting. It describes how some of the more traditional resources can reveal a great deal about the surf, especially when your favourite surfing website gets it wrong and you don’t know why.

Modern wave forecasts use complex mathematical models in giant supercomputers. They can churn out detailed predictions of all the different wave heights, directions and periods that exist at thousands of points all over the ocean. This is called the **directional spectrum**, and usually we see only a brief summary of that data. Chapter 13 looks a little deeper into the directional spectrum, and how it can tell us so much more about the quality of the waves arriving on our coasts.

An increasing number of surfers live in places where there is just not enough room for the waves they ride to have propagated thousands of miles from some distant storm. Here, the waves are ridden literally inside the storm itself. Chapter 14 describes the peculiarities of surfing in the storm, and reminds us that not everybody has the luxury of clean swells and ruler-edged lines.

The subjects of which this book merely scratches the surface are some of the most conceptually difficult and mathematical: fluid mechanics, dynamical meteorology and physical oceanography. Predicting the size and quality of the waves arriving on the coast is one notch up on the difficulty scale from predicting the weather. Add to that the numerous factors that we, the surfers, consider important, but which have so far been ignored by scientists, and it becomes clear why predicting the surf to any great accuracy is extremely difficult.

Even without trying to predict the surf, just knowing how it got the way it is on any particular day is complicated enough. There are so many contributing factors that go into making the waves the way they are – wind, tide, bathymetry, swell direction, swell quality and a host of other infinitely variable parameters – that accurately describing the surf in all its details is perhaps just as difficult as trying to predict it.

This book can only give the basics about waves and wave forecasting. But hopefully it will answer a few questions, and so enhance your appreciation of waves, surf and the natural environment that surrounds them.

‘What’s the prediction for Saturday,’ asked a third surfer.

‘Flat,’ said the first.

‘Thank goodness for that,’ said the third. ‘I’ve got a wedding on Saturday, and I can’t go surfing.’

‘Whose wedding is it?’ asked the first.
‘Mine,’ said the third.

2 Large-scale Air Movements

Introduction

In this chapter we explore the first stage in the journey of that energy packet I was talking about in Chapter 1. Before a wave is even born, a complex series of processes takes place between the Sun's incoming radiation, the formation of weather systems, and the generation of waves on the ocean. Through a series of progressively more realistic models of the Earth, we arrive at a global circulation pattern – a framework which forms the basis behind our day-to-day weather, including the wind and waves. In this chapter we also look at the **Coriolis force** – a fundamental mechanism in oceanography and meteorology. The Coriolis force makes low and high pressures rotate the way they do; it affects ocean currents as they circulate around the globe, and it makes tides rotate around imaginary axes called **amphidromic points**.

The global circulation

A great deal of what happens on Earth is directly or indirectly driven by the Sun's energy. The Sun is like a huge battery, supplying us with a universal source of power through its seemingly inexhaustible process of nuclear fusion.

Not surprisingly, the energy needed to supply our weather comes from the Sun. The uneven heating of the Earth's surface by the Sun is the first link in a chain of events leading to tornadoes, thunder storms, blizzards and the ocean-borne depressions ultimately responsible for the waves we ride.

The fundamental cause of our weather is the fact that the Sun's energy does not heat up the poles and Equator evenly. The Equator is hotter than the poles, and the atmosphere tries to counteract this by continually attempting to redistribute the heat evenly over the Earth's surface. This makes the air move around in average circulation patterns which meteorologists call **statistical highs** and

statistical lows. These are a good starting point before we start to explore the more complex patterns that complicate our weather on a day-to-day basis.

In winter, it is normal to find deep low-pressure systems over the sea and strong high-pressure systems over the land, whereas in summer we typically see large high-pressure systems over the oceans and weak low-pressure systems over the land. How does this come about? What happens to the atmosphere after the Earth's surface has initially been heated up by the Sun's rays?

To investigate this, we must take things step by step – starting with a very simple and highly unrealistic model of the Earth, and then adding various factors to make it progressively more complicated and realistic.

First, we will consider a fictitious Earth, totally covered in water, with no rotation and no seasons. We will see what happens to the atmosphere when it is heated by the Sun. Then we will add the effects of (a) the Earth's rotation about its own axis, (b) the Earth's rotation around the Sun, and (c) the presence of the continents.

A landless, stationary Earth

In this, the most simple model, the Earth is completely covered in water, does not rotate, and there are no seasons.

The Equator is hotter than the poles because it receives a larger amount of solar radiation. This is because the Sun's rays strike the surface of the Earth almost straight downwards at the Equator but at a highly oblique angle at the poles. As a result, the same amount of incoming energy is spread over a larger area at the poles, which makes it less concentrated.

Because the Earth's surface at the Equator is relatively hot, the surface air expands and becomes less dense than its surroundings. This means that it will rise through the process of **convection** and leave a large gap, allowing air to come in and fill that gap. So, on the Earth's surface, air continually flows towards the Equator from north and south. The air that rises at the Equator, once it reaches the upper atmosphere, is forced to travel north and south towards the poles, where it sinks to the surface again. A giant, three-dimensional circulation pattern is set up in each hemisphere: air flowing from the poles to the Equator on the surface, rising at the Equator, travelling back towards the poles in the upper atmosphere, and then sinking again at the poles (Figure 2.1).

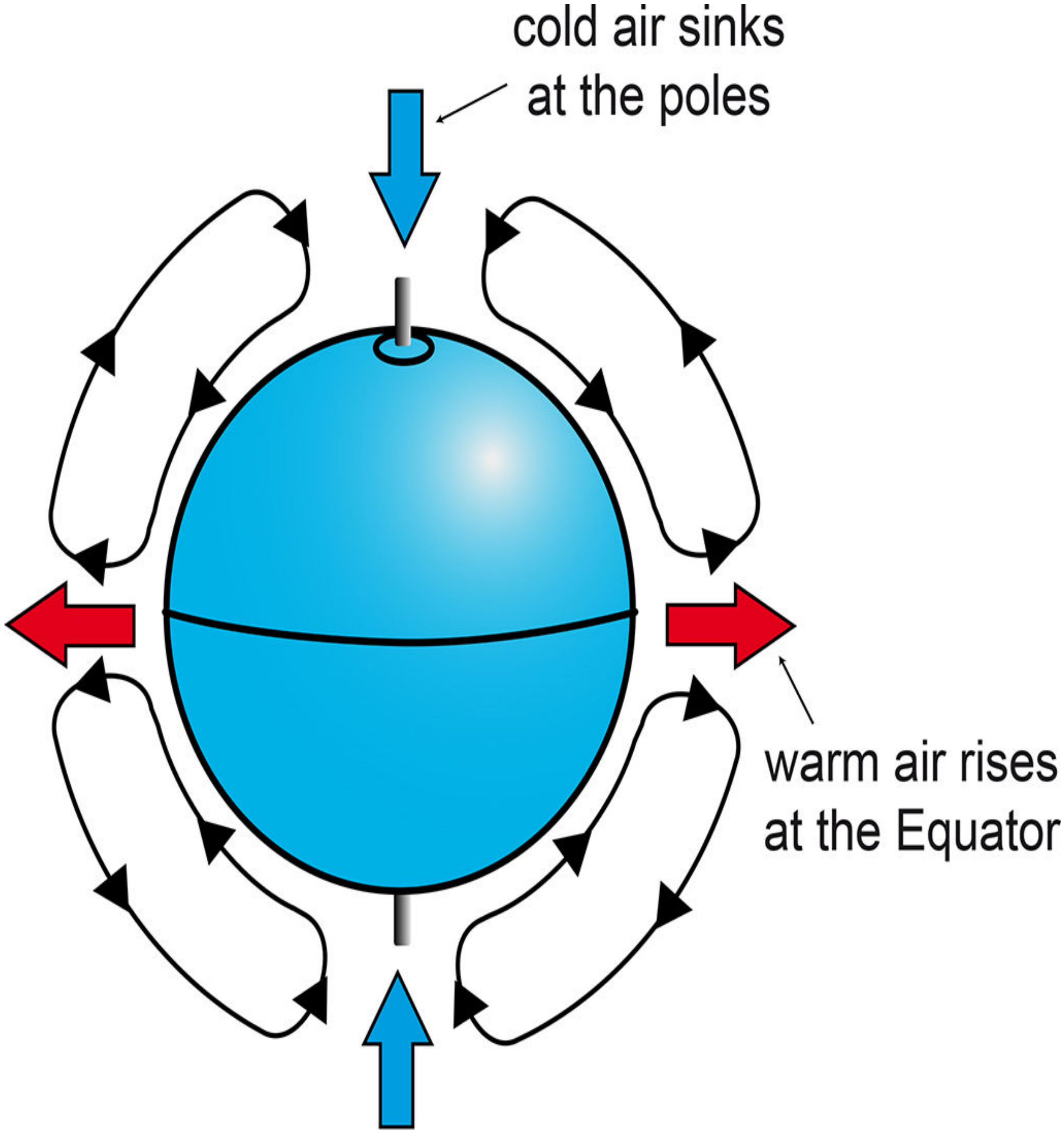

Figure 2.1: The large-scale circulation patterns that would occur on a landless, stationary Earth.

A landless, rotating Earth

Now, in the next model, we will be more realistic and consider the rotation of the Earth about its own axis. The Coriolis force (see later on in this chapter) takes effect and all large-scale movement is diverted to the right in the Northern hemisphere, and to the left in the Southern hemisphere.

Instead of going all the way from the poles to the Equator in a straight line, the surface air is side-tracked – it twists back on itself about a third of the way between the poles and the Equator. The effect of the Coriolis force causes the large circulation cells mentioned above to be ‘short-circuited’ – the result being six spiralling wind-bands, three in each hemisphere.

The air at around latitude 60° north and south is continually rising, spiralling upwards. This constant sucking of air away from the surface locally reduces the surface pressure, which helps to explain why the formation of low-pressure systems is favourable at latitudes between 40° and 70°.

Conversely, the air around 30° north and south is persistently sinking, all the time trying to pump more air into the surface layers, which increases the surface pressure. This is why semi-permanent high-pressure systems like the Azores High or the North Pacific High tend to be found at about 30° latitude.

We now have a rotating, water-covered planet with alternate low- and high-pressure belts around its surface, and a three-dimensional spiralling circulation pattern. The general position for the formation of lows and highs has more or less been established, and the existence of the Coriolis force ensures that the lows and highs are circulating the way they are supposed to (Figure 2.2).

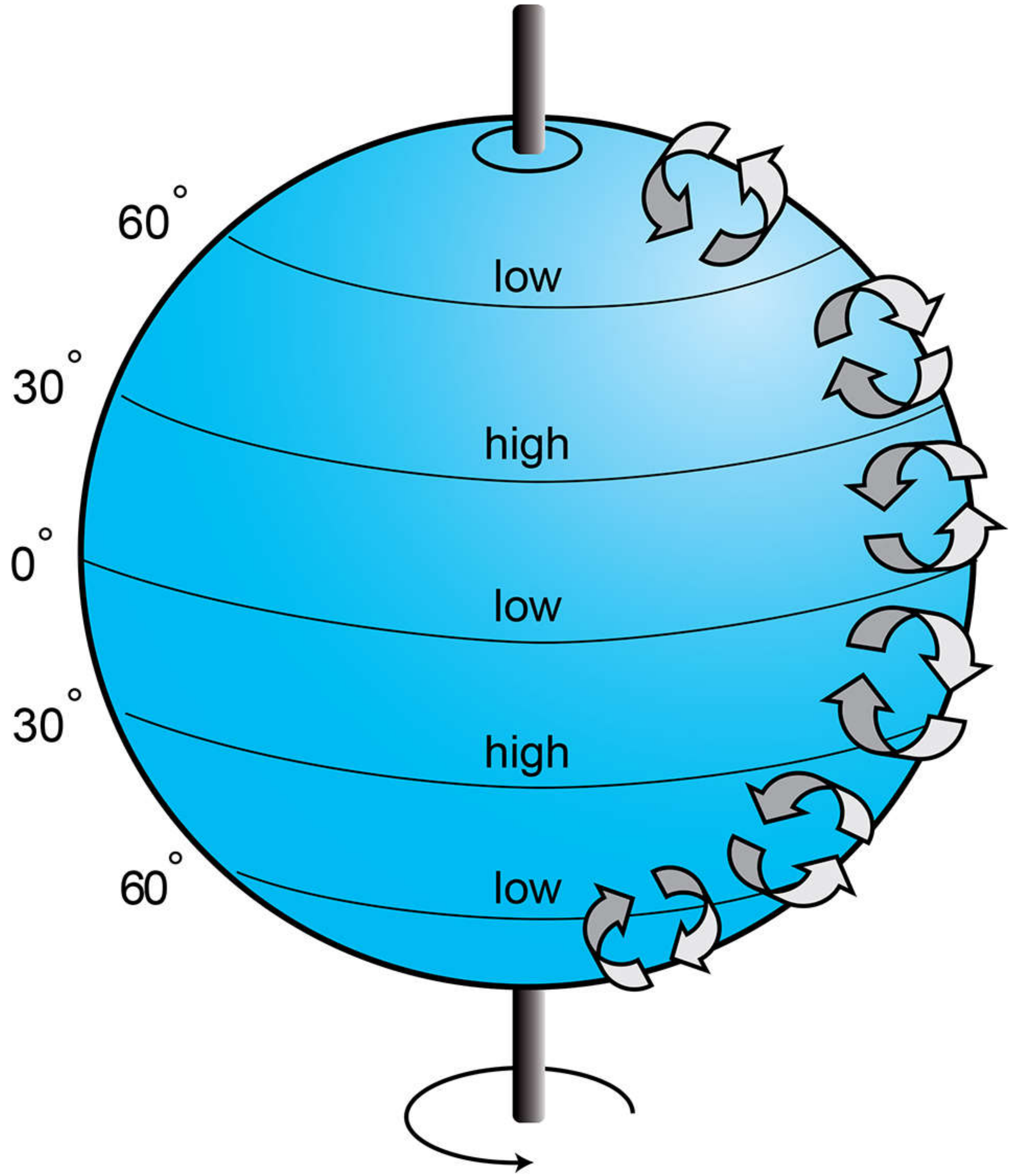

Figure 2.2: Spiralling wind patterns that would result on a landless, rotating Earth.

A landless, rotating Earth with seasons

So far, we have a planet covered entirely with water that is rotating about its own axis once every 24 hours. We have said nothing about its rotation around the Sun every 365 days. If the Earth's axis was sticking straight up and down, then the Earth's rotation around the Sun would not make any difference to our simple model. What complicates matters is the fact that the Earth rotates around the Sun inclined at an angle of 23.5° – an angle called the **obliquity of the ecliptic**. Depending on which pole is closer to the Sun, one hemisphere is warmer than the other at any given time of the year, again due to the angle with which the Sun's rays strike the surface of the Earth. (Note that, at the vernal and autumnal equinoxes – 21 March and 21 September respectively – both hemispheres briefly receive the same amount of sunshine.)

The angle presented to the incoming rays by the surface of the Earth, and therefore the amount of heat received, varies more throughout the year at the poles than it does at the Equator. This causes much greater seasonal variations in polar temperatures than Equatorial ones. In February, for example, the North Pole is much colder than the Equator. In August, however, the North Pole is quite warm, but the Equator is still virtually the same temperature as it was in February. So, in winter, the difference between polar and Equatorial temperatures is very large; but in summer, the difference between polar and Equatorial temperatures is quite small (Figure 2.3).

Thinking back to our original idea of air movement due to uneven heating, we can see that this air movement is massively enhanced during the winter months due to the extra temperature difference. Those spiralling bands work overtime to give us more wind, deeper lows and bigger surf in the winter.

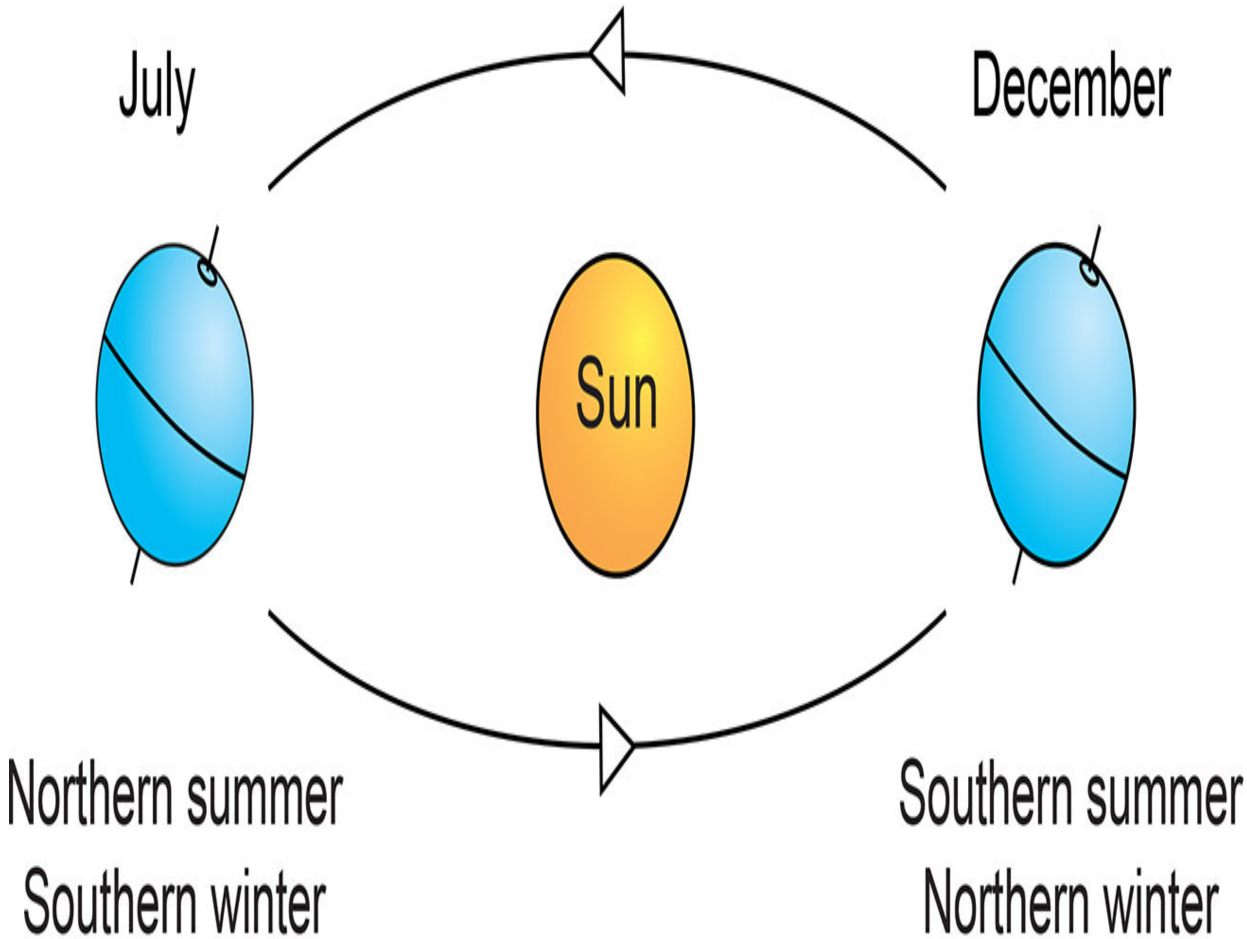

Figure 2.3: The inclination of the Earth relative to the Sun in July and December.

A rotating Earth with land, sea and seasons

The previous models have considered a fictitious planet completely devoid of all land. This next one considers an Earth containing continents.

The first thing to realize is that land and water respond differently when they heat up and cool down. Water has a much greater

specific heat capacity (**SHC**) than land – in other words, you have to pump more energy into the same amount of water to heat it up by the same amount as you do with land. Therefore, water takes a lot longer to heat up and cool down than land does. While the temperature of the continents can swing many tens of degrees from winter to summer, the oceans are content to keep a more constant average temperature throughout the year. (More about this in Chapter 8.)

So summer means warm land and not-so-warm oceans, and winter means cold land and not-so-cold oceans. Broadly speaking, in the summer the land is warmer than the sea, and in the winter the sea is warmer than the land. As a result, there is more convection over the land than over the sea in summer, and there is more convection over the sea than over the land in winter. Ultimately, this leads to a general summer pattern of relatively low pressure over the land and relatively high pressure over the sea, and a general winter pattern of relatively high pressure over the land and relatively low pressure over the sea. In the new model, the six-band system of high and low pressure shown in Figure 2.2 is broken up in an east-west direction, with different patterns for summer and winter.

Don’t forget, this is just the average situation; it does not explain the daily, weekly or monthly variations in our weather patterns. But it does go some way to explaining typical average summer and winter patterns: namely, a large high pressure over the sea and thundery low pressures over the land in the summer, and a deep low pressure over the sea and a stagnant high pressure over the land in the winter.

The last important thing to add is the fact that there is more land in the Northern hemisphere than there is in the Southern hemisphere. The east-west pressure variation due to the presence of land, as explained above, is much more noticeable in the Northern hemisphere than in the Southern hemisphere, simply because the Northern hemisphere contains more land. The seasonal variations in the Southern hemisphere are considerably weaker, and its circulation patterns are much simpler. A major feature of the Southern hemisphere is the **roaring forties** – a band of westerly winds that blow continuously around the globe between 40° and 60° degrees south, and a series of high pressures just to the north. In summer, the whole pattern shifts south and the westerlies weaken; in winter, the whole pattern shifts north and the westerlies strengthen (Figure 2.4 and Figure 2.5).

The extra simplicity and lack of seasonal variation in the Southern hemisphere mean that wave heights tend to be more evenly spread throughout the year than they are in the Northern hemisphere. The difference between average summer wave heights and average winter ones on, for example, the west coast of South America or Australia is typically much less than on the west coast of Europe or North America.

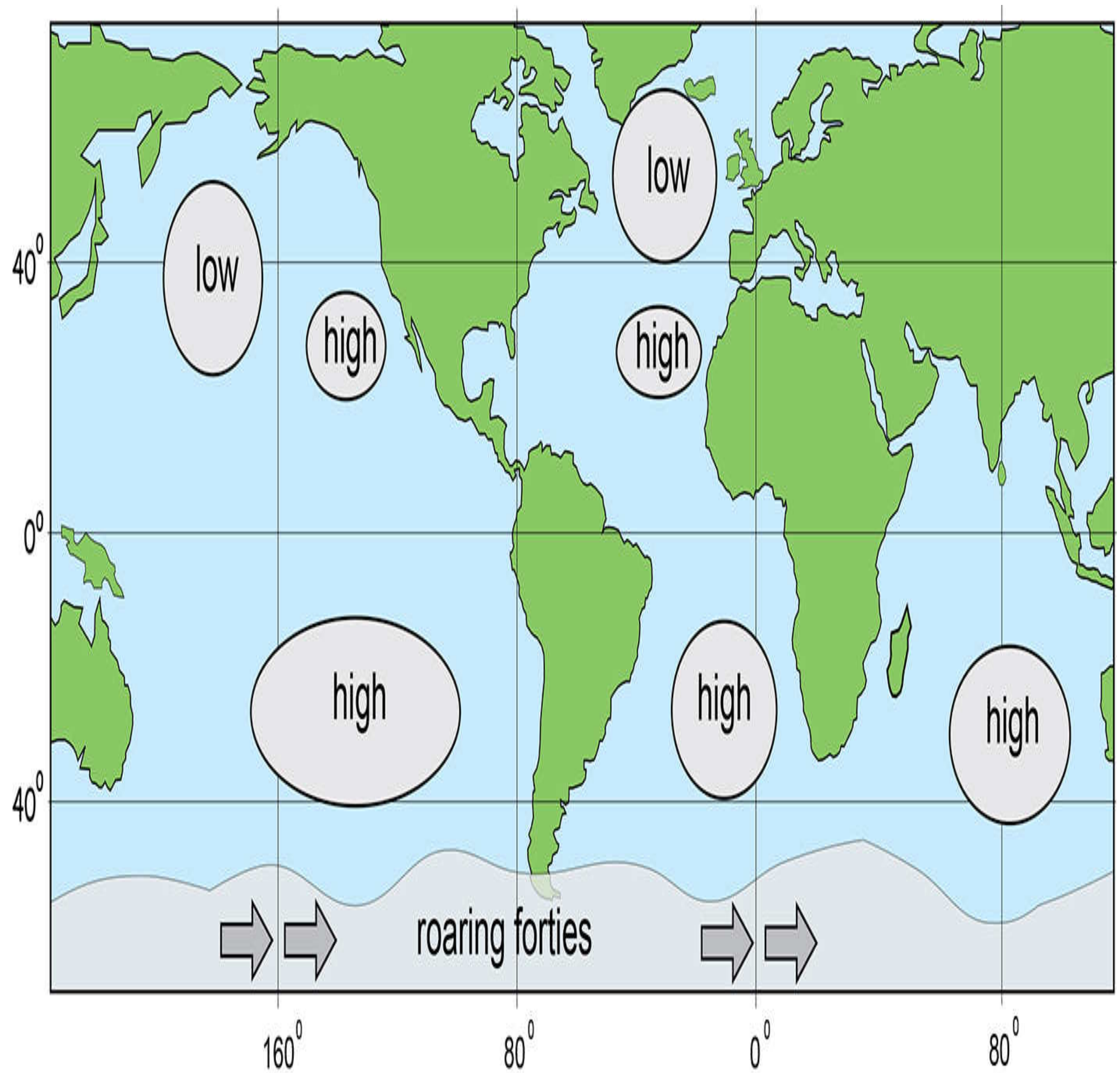

Figure 2.4: Simplified illustration of 'averaged' weather patterns over the oceans for January.

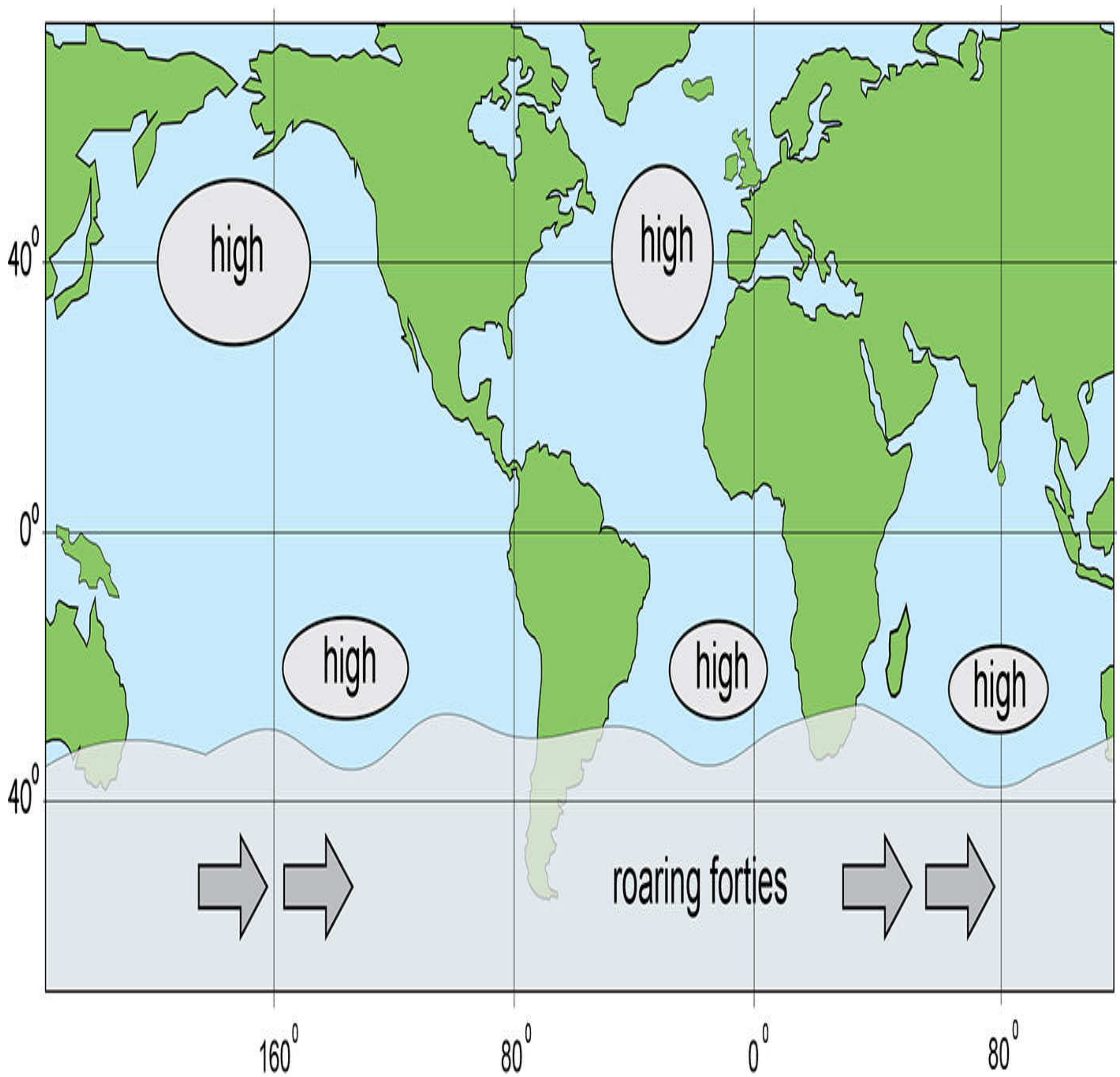

Figure 2.5: Simplified illustration of 'averaged' weather patterns over the oceans for July.

The Coriolis force

The Coriolis force crops up in oceanography and meteorology time after time. A basic appreciation of this strange but important phenomenon is essential if we are going to talk about such things as low-pressure and high-pressure systems, oceanic currents and tides. In fact, apart from very small-scale processes, almost everything in the world of physical oceanography is affected by the Coriolis force.

The Coriolis force was discovered in the early nineteenth century by the French physicist Gustave Gaspard Coriolis. It is the reason why all large-scale motions turn right in the Northern hemisphere and left in the Southern hemisphere. It is the reason why lows and highs, ocean currents and tides circulate the way they do.

The Coriolis force is sometimes referred to by scientists as an 'apparent' force, not a 'real' one. This is because the object in question (air in the atmosphere, or water in the ocean) only turns to the right or left from the point of view of an observer on the surface of the Earth. If you were in space, not rotating with the Earth around its own axis, you would observe the object to travel in a straight line. The following rather precise definition captures this in a nutshell:

> *'The tendency for water movement to maintain a uniform direction in absolute space means that it performs a curved path in the rotating frame of reference within which our observations are made on Earth.'* (David Pugh, *Changing Sea Levels: effects of tides, weather and climate*)

The full explanation of the Coriolis force is highly mathematical, and does not belong in this book. However, there are several ways of explaining it in simple terms, one of which is the following:

The Earth rotates on its own axis from west to east, and takes about 24 hours to complete one cycle. If you look at the lines of latitude, you will see that the biggest one is the Equator itself, and that they become smaller towards the poles. If you were on the Equator you would be covering more distance in those 24 hours than you would, say, in London. If you were right on one of the poles you would not be covering any distance at all – you would just be going round in the same spot. So, the surface of the Earth is travelling very fast in an easterly direction at the Equator (about 1,600 km/h), and progressively more slowly at latitudes nearer the poles. The eastward speed of a point 60° north, for example, is about 800 km/h.

To begin to understand the Coriolis force, think in terms of a single ‘air parcel’, similar to a large balloon. This air parcel is embedded in the rest of the atmosphere, which is rotating with the Earth’s surface. The air parcel is fixed to the surrounding atmosphere at first, but we can ‘unstick’ it and move it around. What happens to this air parcel as we try to move it around is the key to the Coriolis force.

Think of the air parcel in London. Before we do anything to it, it is already travelling from west to east at a certain speed – let’s call it ‘London’ speed. If we suddenly push it southwards towards Madrid, it will still be going from east to west at ‘London’ speed, but the ground beneath it is now going at ‘Madrid’ speed, which is faster than ‘London’ speed. Therefore, the air parcel will lag behind relative to the Earth’s surface. Instead of going directly north-south, it will veer off to the right (Figure 2.6).

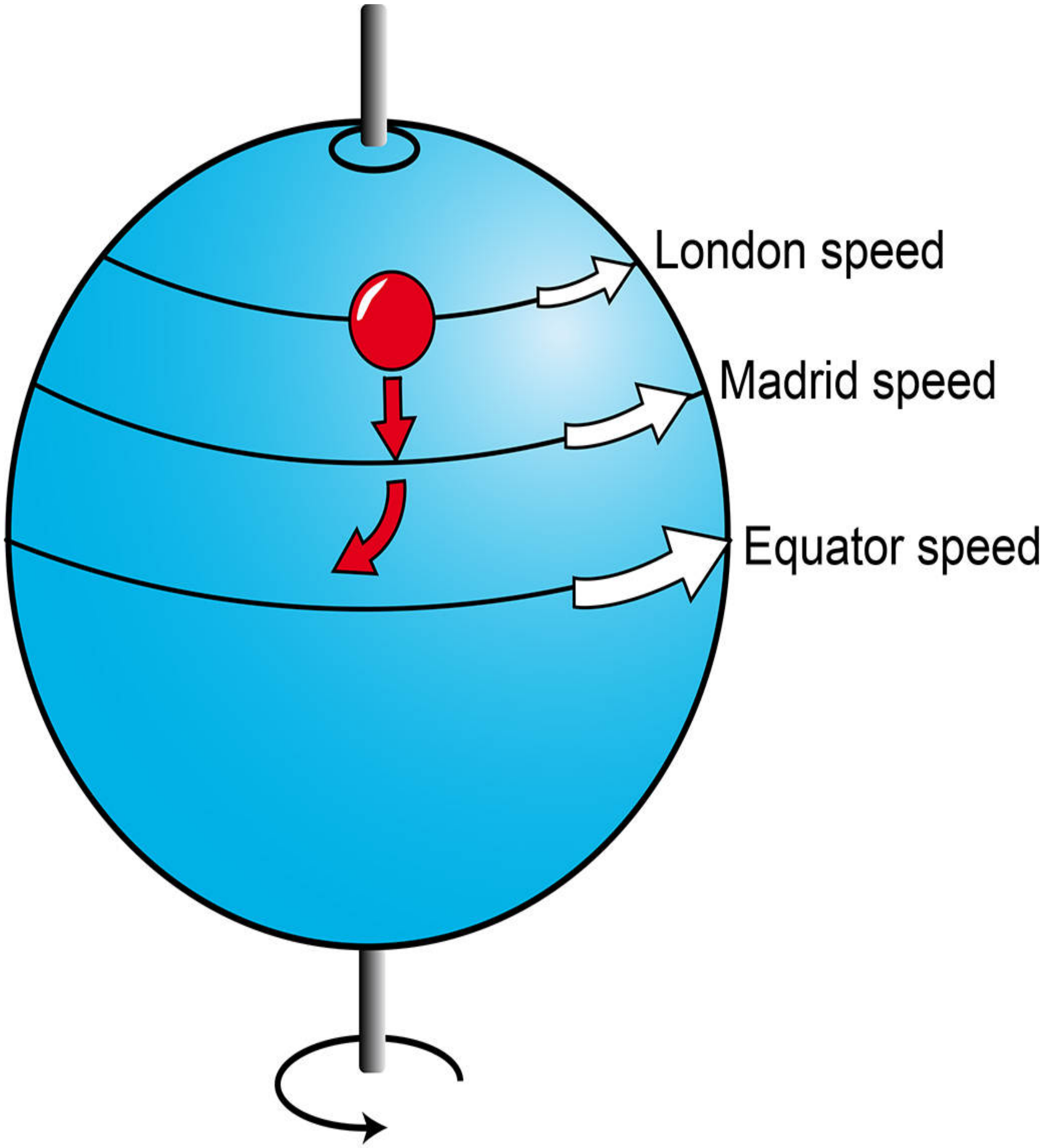

Figure 2.6: An air parcel forced south will end up where the eastward velocity of the Earth is faster, so will swing to the right.

However, if the air parcel starts off in Madrid, and we push it northwards towards London, the opposite will happen. The air parcel, having carried with it its initial ‘Madrid’ west-east velocity, will find itself travelling eastwards faster than the Earth beneath it, and so tend to swerve to the right.

Now, what happens to the air parcel if we force it in an east-west direction? Say the air parcel is in London again, travelling from west to east with the Earth at ‘London’ speed. This time we give it a push eastwards towards Moscow. What we have effectively done is increase its rotational speed slightly. This increase in speed tends to throw the air parcel outwards, which causes it to increase its radius of curvature and move to a position where the Earth is fatter. So the air parcel, instead of going east towards Moscow, swerves south towards the Equator. In other words, it turns to the right. Conversely, if we give the air parcel a westward push from Moscow towards London, we effectively decrease its rotational speed. This will make it want to travel around a smaller radius of curvature, nearer the North Pole. Instead of ending up in London, it veers off towards Scotland; again, it veers to the right (Figure 2.7).

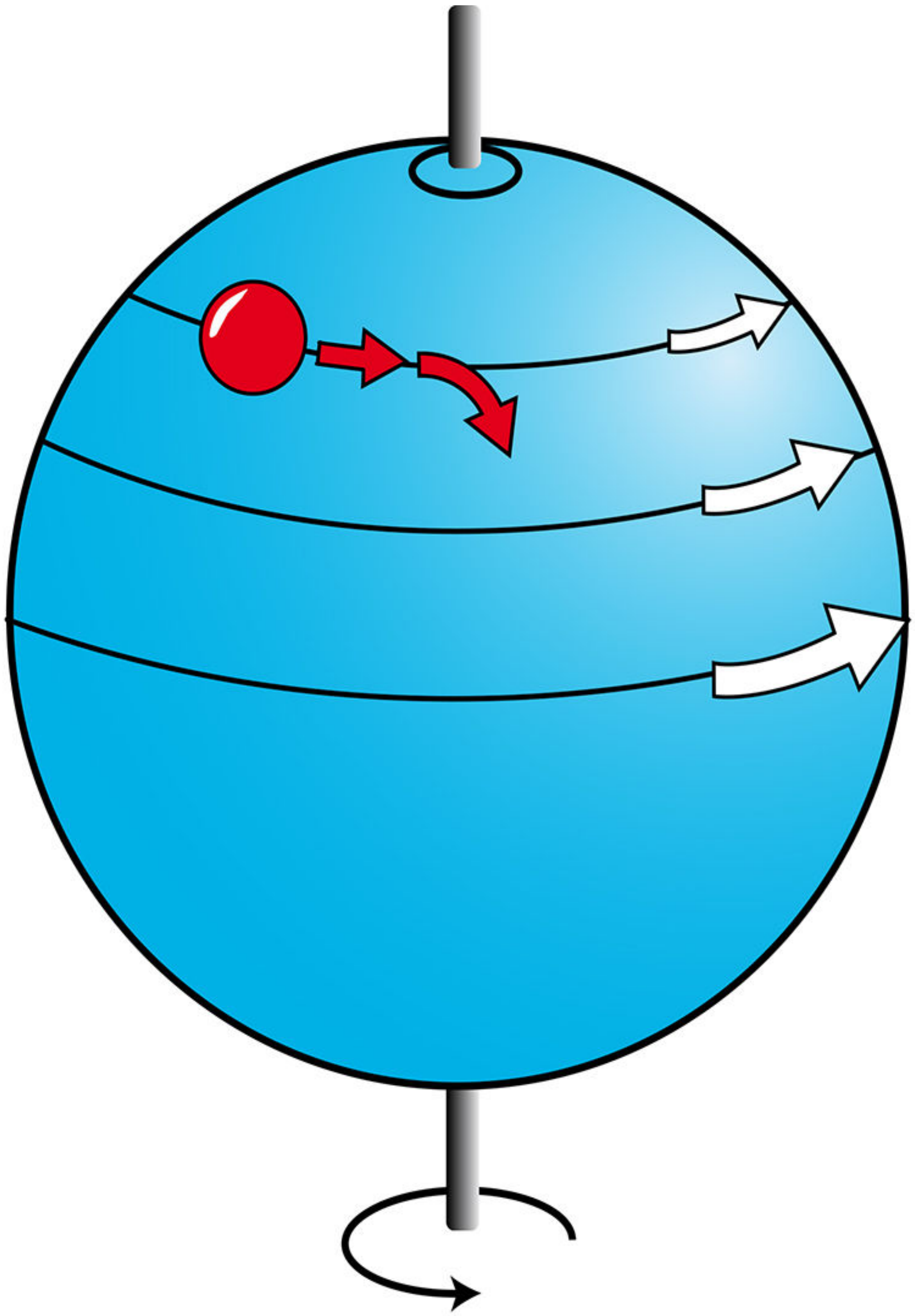

Figure 2.7: An air parcel pushed east will try to increase its radius of curvature, so it will swing to the right.

In summary, whichever way we push the air parcel, it swerves to the right on the surface of the rotating Earth. In the Southern hemisphere, the whole system is mirror-imaged. By exactly the same principles, the air parcel will always be deflected to the left.

How does the ‘air parcel’ story help us to understand why low-pressure and high-pressure systems rotate the way they do? First, we must look at weather systems as if they were made up of an infinite number of discrete air parcels. The natural tendency for one of these is to try to travel from high pressure to low pressure.

Imagine a cell of low pressure in the Northern hemisphere, surrounded by air at a relatively higher pressure. Immediately, a stream of air parcels tries to rush into the low pressure from outside – trying to ‘fill up’ the low and equalize the pressure. But they get deflected to the right by the Coriolis force, veer off to the right, and end up circulating around the low pressure in an anticlockwise direction.

The same principle can be applied to a high-pressure cell surrounded by relatively low pressure. The air tries to flow outwards from the high, but is deflected to the right and ends up circulating around the system in a clockwise direction. Again, this can easily be mirrored into the Southern hemisphere, where the air circulates around low pressures in a clockwise direction, and around high pressures in an anticlockwise direction (Figure 2.8).

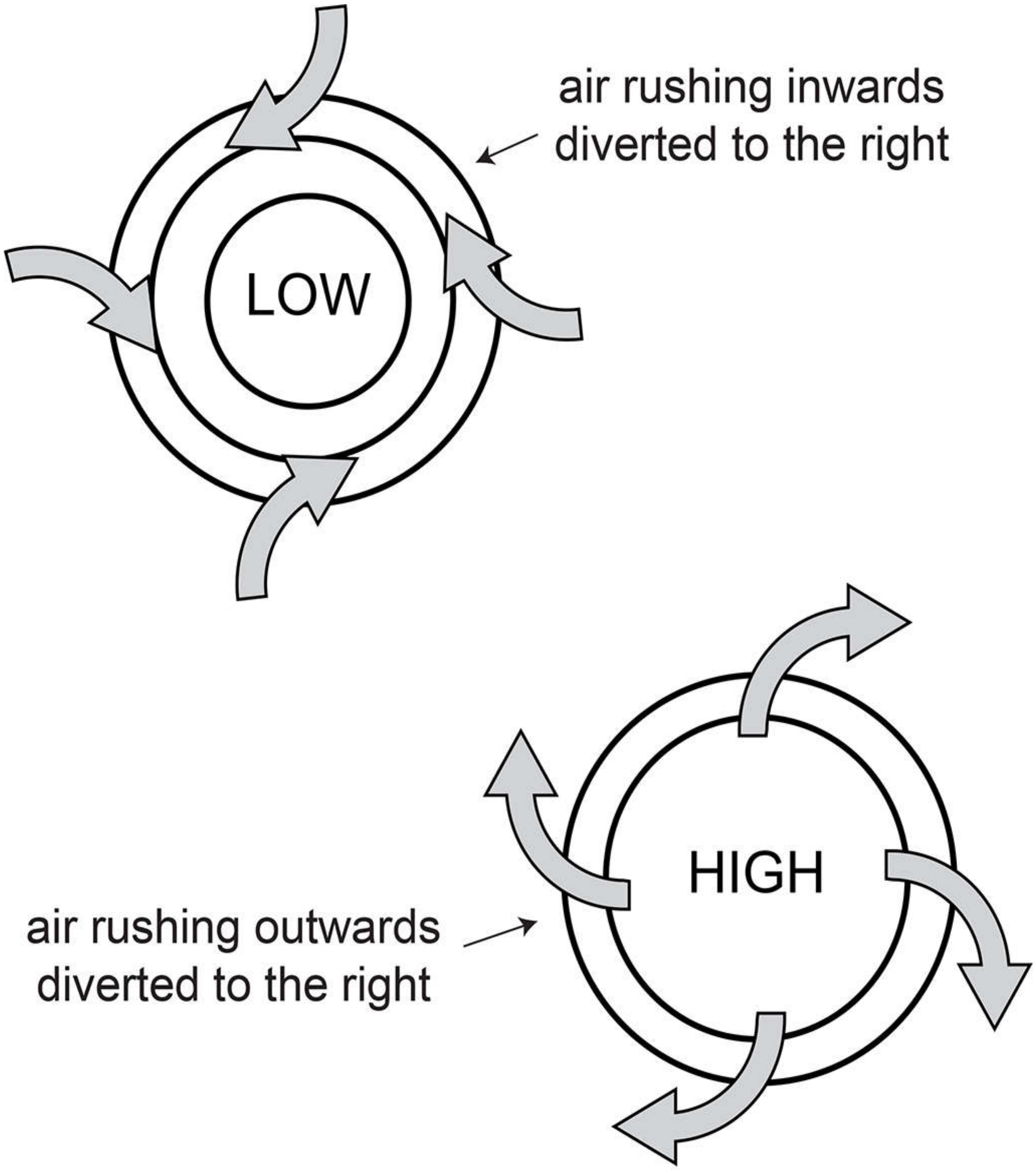

Figure 2.8: A Northern hemisphere low and high showing how the Coriolis force makes the air circulate around each system.

Note that the term **cyclonic** means the direction of rotation of a **cyclone** – essentially a low pressure system – independent of which hemisphere you are talking about. Cyclonic rotation means anticlockwise rotation in the Northern hemisphere and clockwise rotation in the Southern hemisphere. The term **anticyclonic** is, of course, the direction of rotation of an area of high pressure, or **anticyclone** – clockwise in the Northern hemisphere and anticlockwise in the Southern hemisphere.

3 Formation of a Depression

Introduction

Without doubt, the most important phenomenon for producing the waves we ride is the low pressure, also called the **mid-latitude depression** or **extra-tropical cyclone**. The low pressure is really just a cell of air whose pressure is lower than its surroundings. However, thanks to the Coriolis force, it also features a swirling pattern of fast-moving surface air, which generates waves on the sea by transferring energy from the air to the water. The greater the pressure difference between the centre of the depression and the surrounding air, the faster the air moves. The faster the air moves, the more energy it imparts on the water and the bigger the waves will be.

In this chapter we'll look at how a low-pressure system forms in the first place. The full explanation is a complicated story, actually still not fully understood by scientists. So we'll keep things very simple.

Beginnings

Some of the first people to study the formation of a low pressure were a group of meteorologists from Bergen in Norway, headed by Vilhelm Bjerknes. They came up with the concept of the **polar front** – a band on the Earth's surface where cold air from the poles meets warm air from the Equator, and where the spawning of the mid-latitude depression takes place. The position of the polar front coincides with that general band of relatively low surface pressure in the large-scale circulation patterns discussed in the last chapter. It is the boundary between two circulation cells, where the surface air is flowing in opposite directions (Figure 3.1).

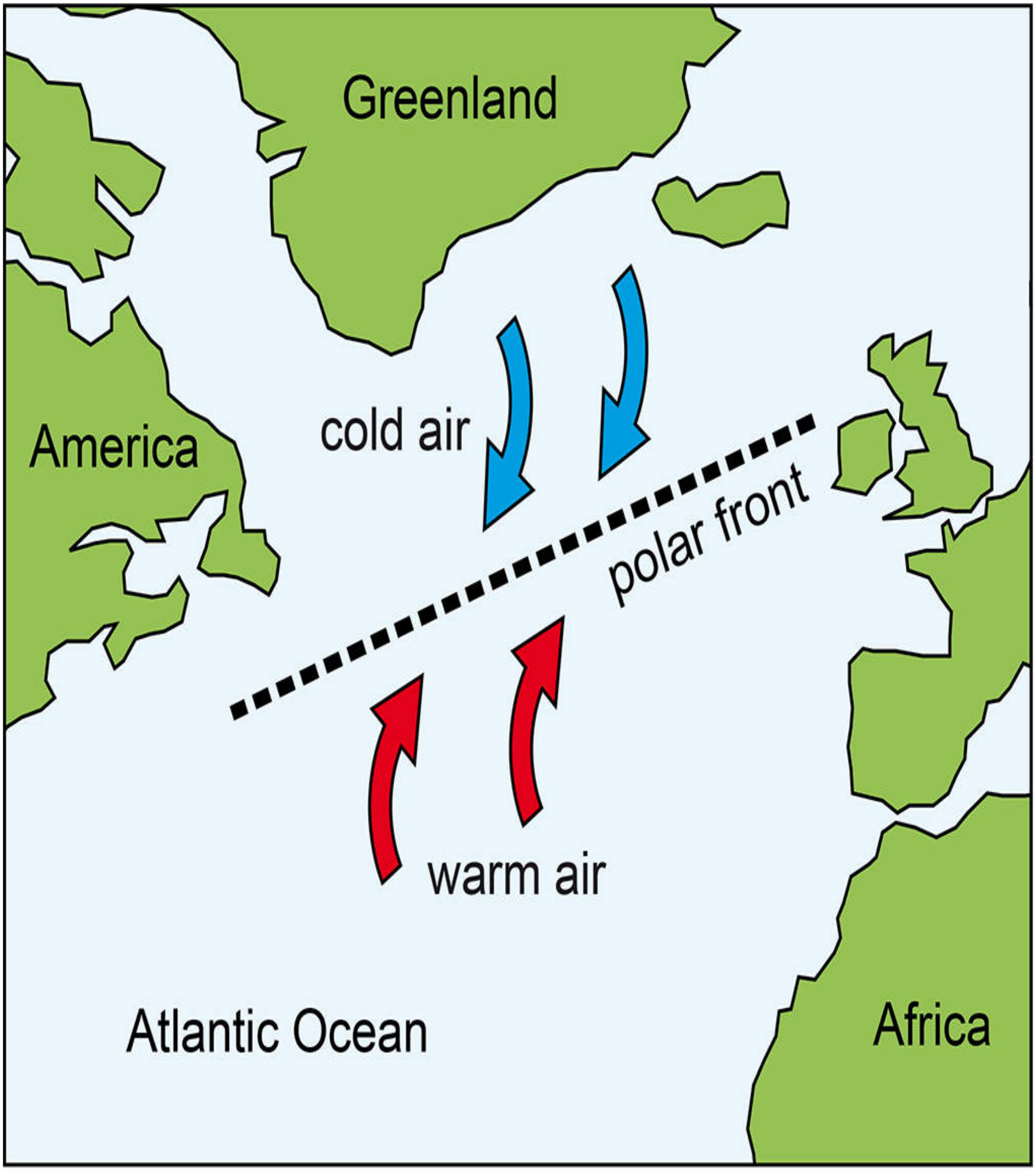

Figure 3.1: North Atlantic polar front, with air masses meeting from the north and south.

In the Northern hemisphere, for example, the polar front is where cold air coming from the north meets warm air coming from the south. Because the warm air is less dense than the cold air it tends to slide over the top of the cold air. This is what is generally happening at the polar front before any kind of low pressure begins to develop.

Through a particular combination of circumstances, a perturbation may appear at some point along the front. For example, the north-south air temperature difference may be particularly intense at this point, or there might be some influence from an external factor like the sea surface temperature. Such a disturbance is known to meteorologists as **baroclinic instability**, meaning that there is a change in pressure (often closely linked to a change in temperature) over a short distance, which causes the atmosphere to become locally unstable. If the perturbation is strong enough and all the factors are right, the front develops a ‘wave’ on it, which grows and intensifies. The process of warm air sliding over the top of the cold air is particularly strong in the area of the perturbation, which leads to a localized drop in surface pressure as the surface air is forced upwards. The newly formed cell of low pressure then begins to suck in air from its surroundings. The air travelling in towards the cell of low pressure is deflected by the Coriolis force, and begins to circulate around the centre of the disturbance in a cyclonic direction (anticlockwise in the Northern hemisphere, clockwise in the Southern hemisphere). This section of the polar front now splits into a system of individual fronts: the **warm front**, behind which is the warm air, and the **cold front**, behind which is the cold air (Figure 3.2).

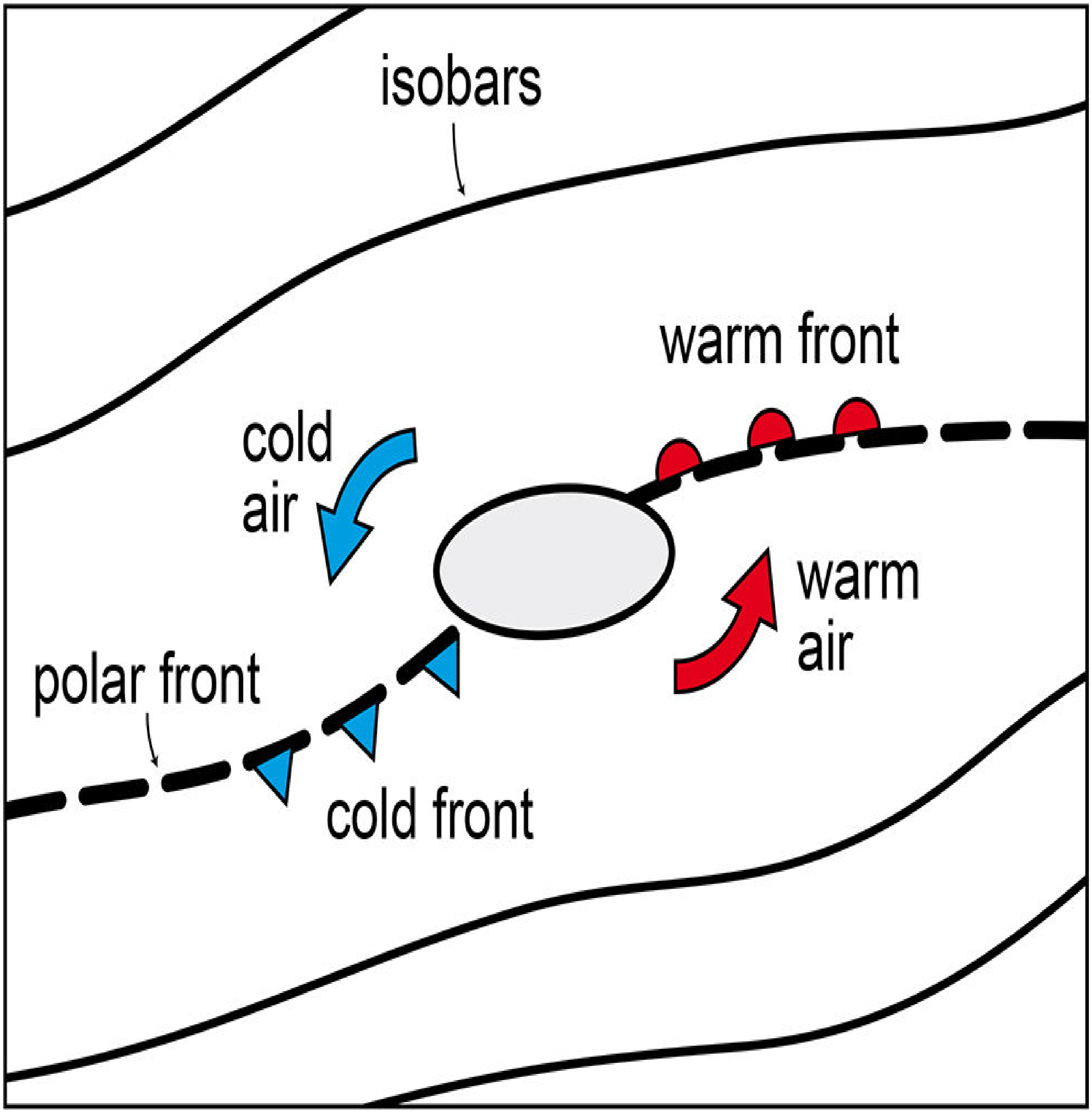

Figure 3.2: Birth of a low-pressure in the Northern hemisphere. Lines marked 'isobars' are contours of equal pressure.

The fully developed low pressure

As the disturbance begins to grow in intensity, the whole thing turns into a large vortex, the air being sucked in from outside and then deflected by the Coriolis force. Between the warm front and the cold front, the **warm sector** develops. The warm sector is an area where the surface winds are strong and blowing in the same direction for some distance – the best conditions for rapid wave growth.

This is the fully developed low-pressure system. If all the factors are right and the surface pressure manages to reach a low enough value, the wind blowing around the system will be strong enough to generate big waves. On the weather chart, a deep low-pressure system can be recognized straight away by a thick mass of closely packed lines of equal pressure, called **isobars** (*iso* = equal; *bar* = pressure'). The closer together they are, the greater the difference in pressure over a given distance – the **pressure gradient**. The greater the pressure gradient, the greater the air flow in towards the centre of the low pressure, and the stronger the winds.

The inrushing air is deflected by the Coriolis force, to the right in the Northern hemisphere, about 75° for a low-pressure system over the ocean. If there were no surface friction, it would be deflected a full 90° and would blow parallel to the isobars. In other words, the winds blow almost parallel to the isobars, but, because of friction, spiral inwards about 15° (Figure 3.3).

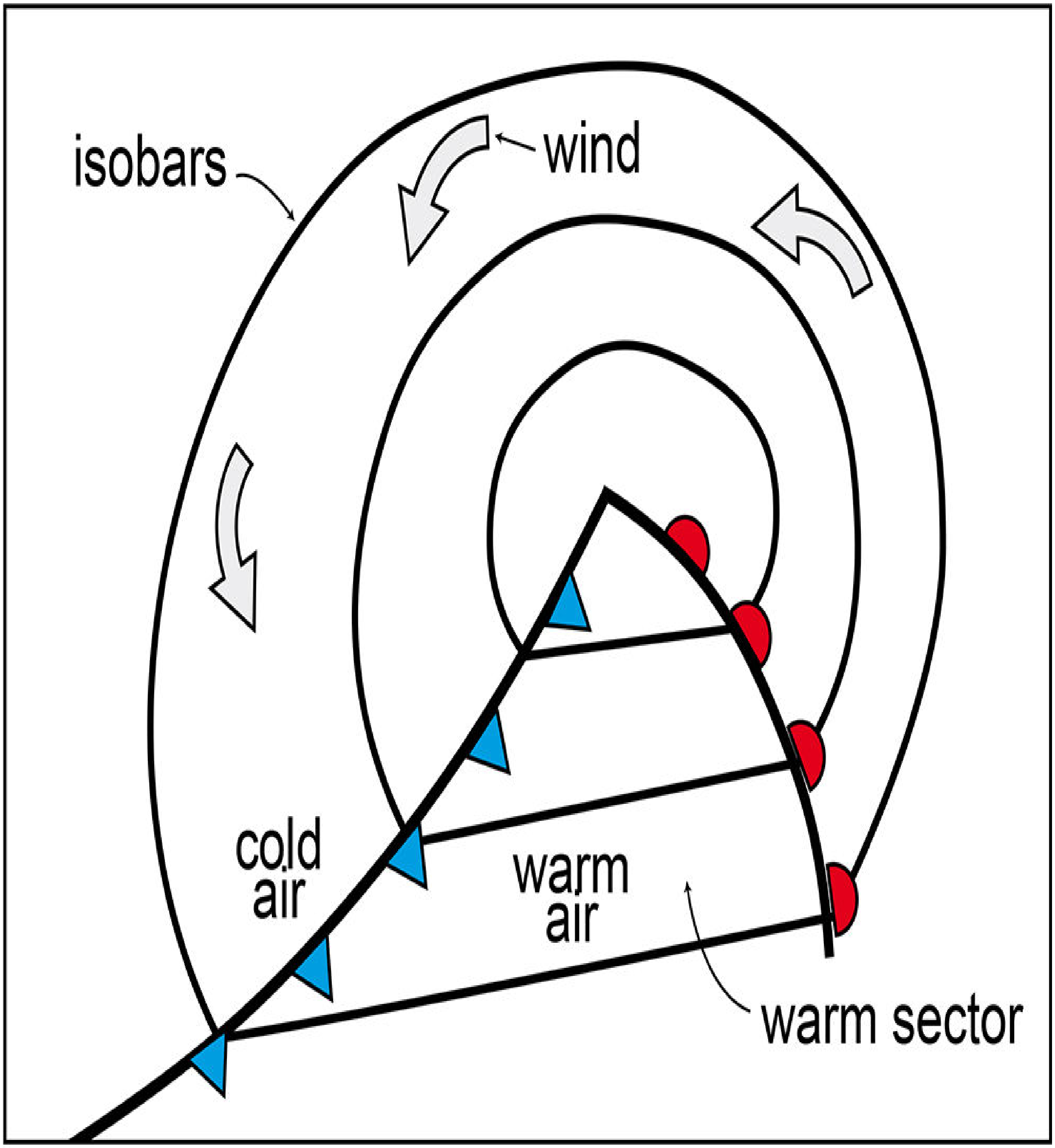

Figure 3.3: Fully developed Northern-hemisphere low-pressure system.

As the low pressure moves from west to east over the ocean, the warm and cold fronts eventually catch up with each other, combining to form an **occluded front**. This is when the whole system starts to weaken. Ultimately, maybe after spawning a few peripheral systems, it loses its identity altogether.

The best situation for generating waves is before this occlusion takes place; when the system is in its fully mature state, with a large warm sector containing strong winds over a long distance. The area of ocean containing the winds that generate a swell is called the **fetch**. The wave height depends on the windspeed, but it also depends on the fetch; the longer the fetch, the bigger the waves. The wave height also depends upon the length of time the wind blows over a particular stretch of ocean – the **duration**. A very small low pressure won’t produce very big waves, even if the winds are quite strong; neither will a system that develops and disappears quickly. A deep low pressure containing a large fetch that stalls in mid-ocean will generate a big swell. Even better is a system that travels along in sync with the swell it is generating, continually pumping energy into the sea surface – a phenomenon called **dynamic fetch**. These are usually the ones that produce the biggest swells of all.

Figure 3.4, Figure 3.5 and Figure 3.6 show a sequence of weather charts from the North Atlantic, 24 hours apart. A clear example can be seen, from the slight disturbance on a frontal system to the fully mature and occluding low pressure, about 48 hours later. The constant broad band of westerly winds on its southern flank ensured huge surf for most of Europe.

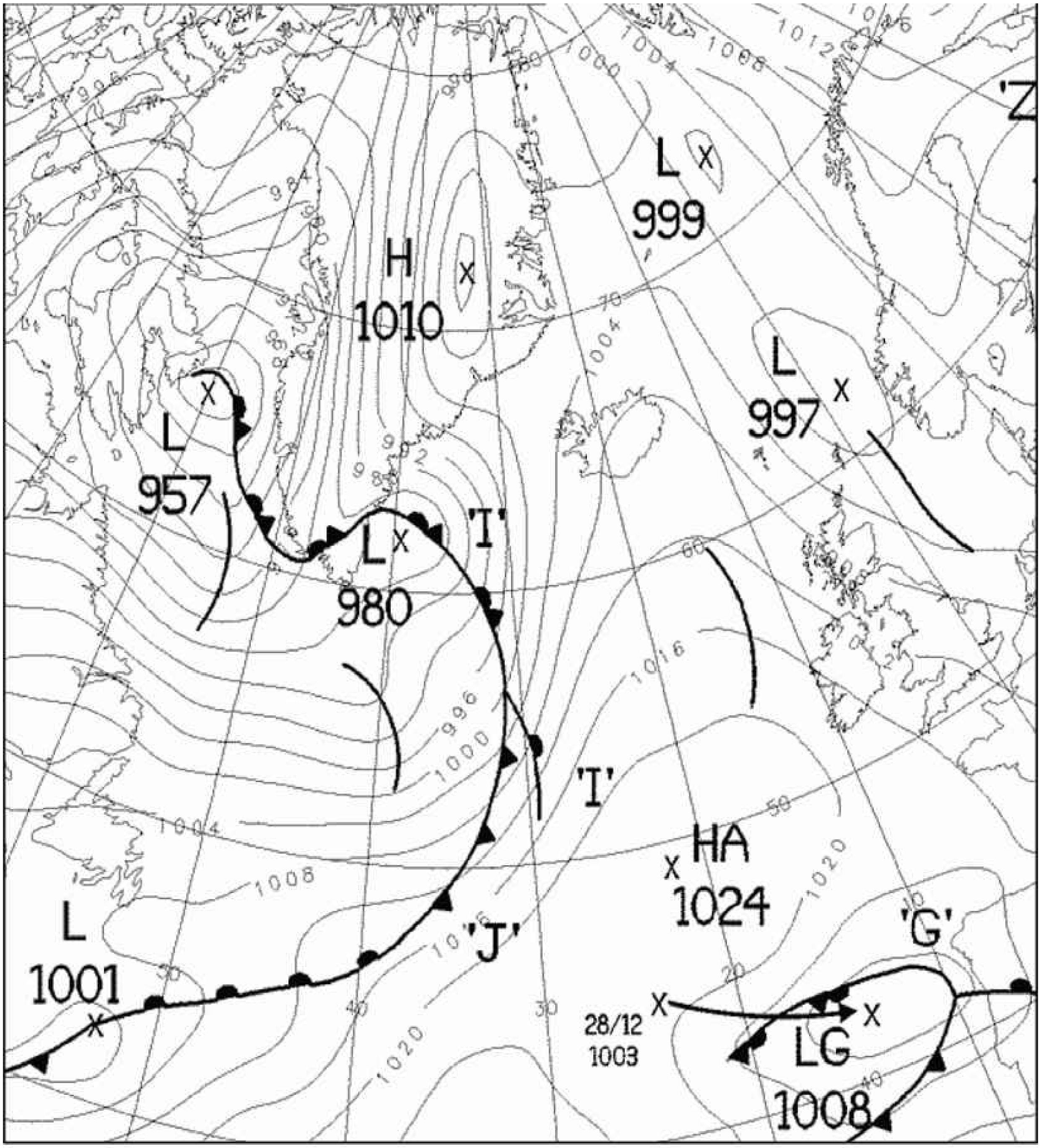

Figure 3.4: The birth of a depression. In the bottom left corner there is a developing low (1001 mb).

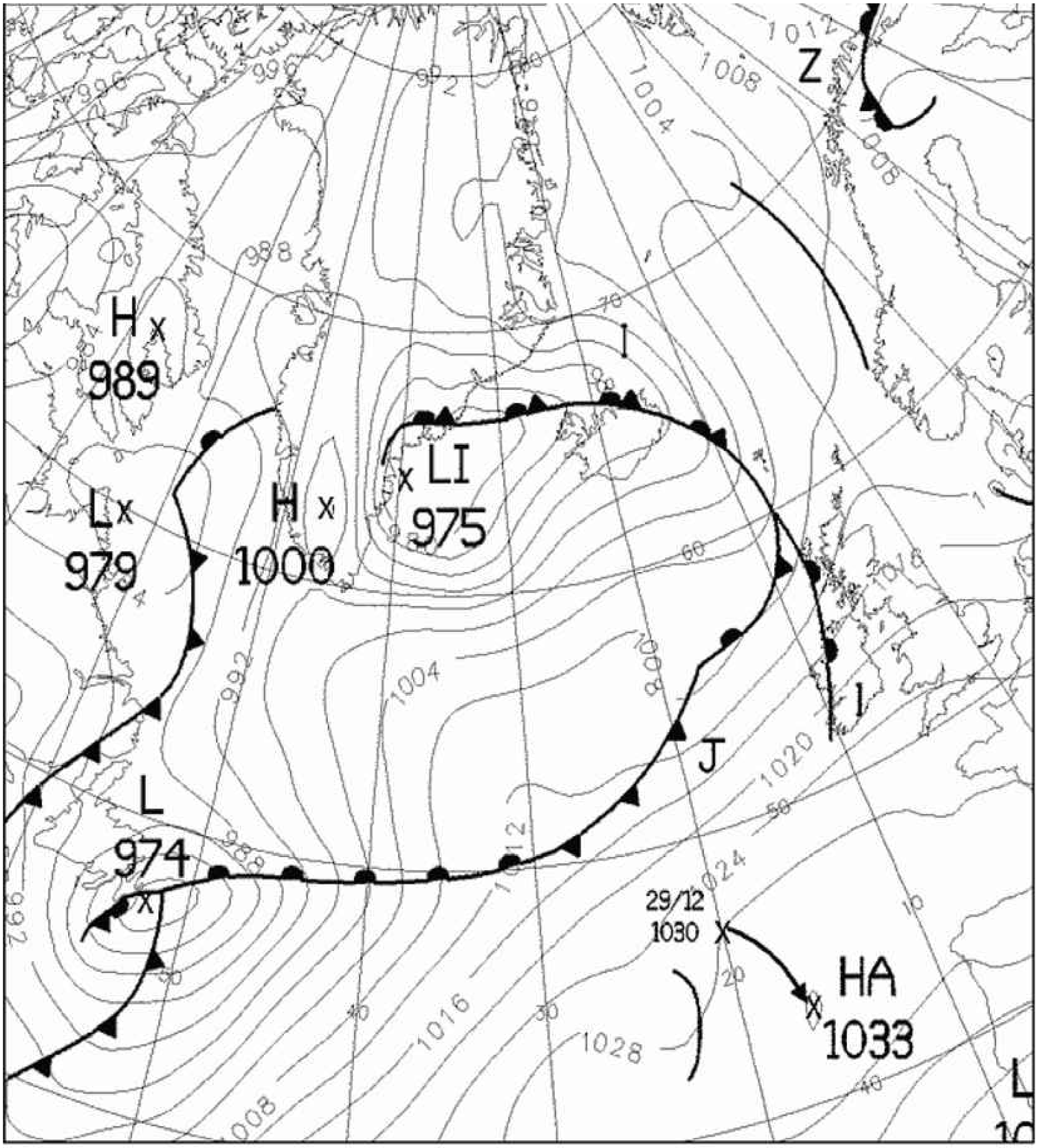

Figure 3.5: Twenty-four hours later, the system has deepened considerably and moved north slightly.

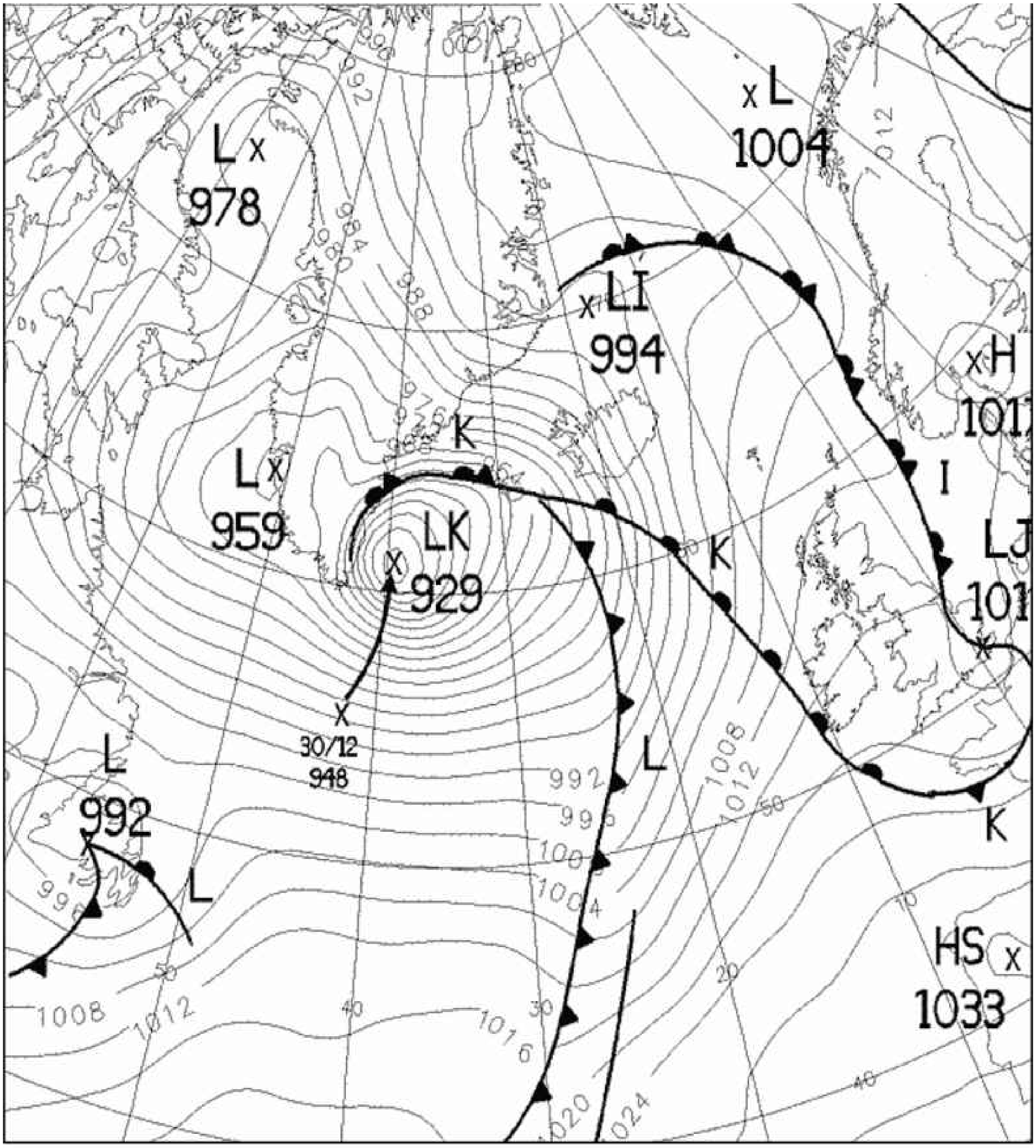

Figure 3.6: Twenty-four hours later again, the low has moved rapidly north-west and deepened to record levels.

Upper air influence

In the charts shown in Figures 3.4 to 3.6 you will have noticed that the low pressure moved extremely quickly north-east as it deepened rapidly. The movement of a low pressure on the surface is highly influenced by the airflow and pressure patterns in the upper atmosphere. Members of the **Bergen school** discovered that the atmosphere high up affects the atmosphere near the surface through a complicated pattern of vertical and horizontal pressure gradients combined with the Coriolis force.

The flow of air in the upper atmosphere tends to be much stronger than on the surface. Some of the strong winds that blow at these altitudes are called the jet stream. The jet stream blows around the globe at latitudes of about 30° to 60°, and at altitudes of about 5,000–10,000 metres above sea level. It goes through phases of blowing fairly straight, but it also tends to have meanders in it, making the flow waver north and south. The position and size of these meanders vary constantly, greatly influencing the formation and trajectory of the surface low-pressure systems.

One of the discoveries made by the Bergen school was that the path taken by a surface depression closely follows the track of the jet stream. The amount of energy pumped into the surface systems by the upper air stream depends on the strength and orientation of the jet stream. The deepest low pressures tend to develop in places where the jet stream is very strong, and where the initial disturbance takes place on the poleward arm of a meander. For example, in the Atlantic, the formation of strong low-pressure systems is most favourable when the jet stream is flowing diagonally from south-west to north-east across the ocean. They tend to form somewhere off the eastern seaboard of North America, and then deepen rapidly as they track towards the British Isles. If the jet stream is particularly weak, split into two paths or orientated from north-west to south-east across the ocean, the formation of strong low-pressure systems is much less likely.

4 Growth of Waves on the Ocean

Introduction

In Chapter 3, we explained the formation of an ocean-borne mid-latitude depression, whose winds are the most common source of swell for the waves we ride. The waves themselves are generated by nothing more than the action of the wind blowing across the surface of the sea, causing energy to be transferred from the air to the water. Exactly how this process takes place is a little more complex than it first sounds and, again, is still not fully understood. In this chapter we will look briefly at the theories behind wind-wave generation currently accepted by scientists and used in wave-prediction models.

The anatomy of a wave

Before we get into wave generation, let me just explain a few wave terms. This is the first time in the book we'll be talking specifically about waves, and from here on you will keep seeing words like *period*, *wavelength*, *amplitude*, *frequency* and *height*. So, just in case you don't know what they mean I will briefly explain them (see also Figure 4.1):

… The **amplitude** is the vertical displacement of the water from its resting position (more precisely, from its average position or the **mean sea level**).
… The **height** is the distance from trough to crest, which is twice the amplitude.
… The **period** is the time taken between when any part of a wave (such as the crest) passes a fixed point and when the same part of the next wave passes that point.
… The **wavelength** is the distance between the same parts of two successive waves. Period and wavelength are only

meaningful when you have a continuation of waves, not a solitary wave.
… The **frequency** is the number of waves that pass a fixed point per second. It is the inverse of the period.

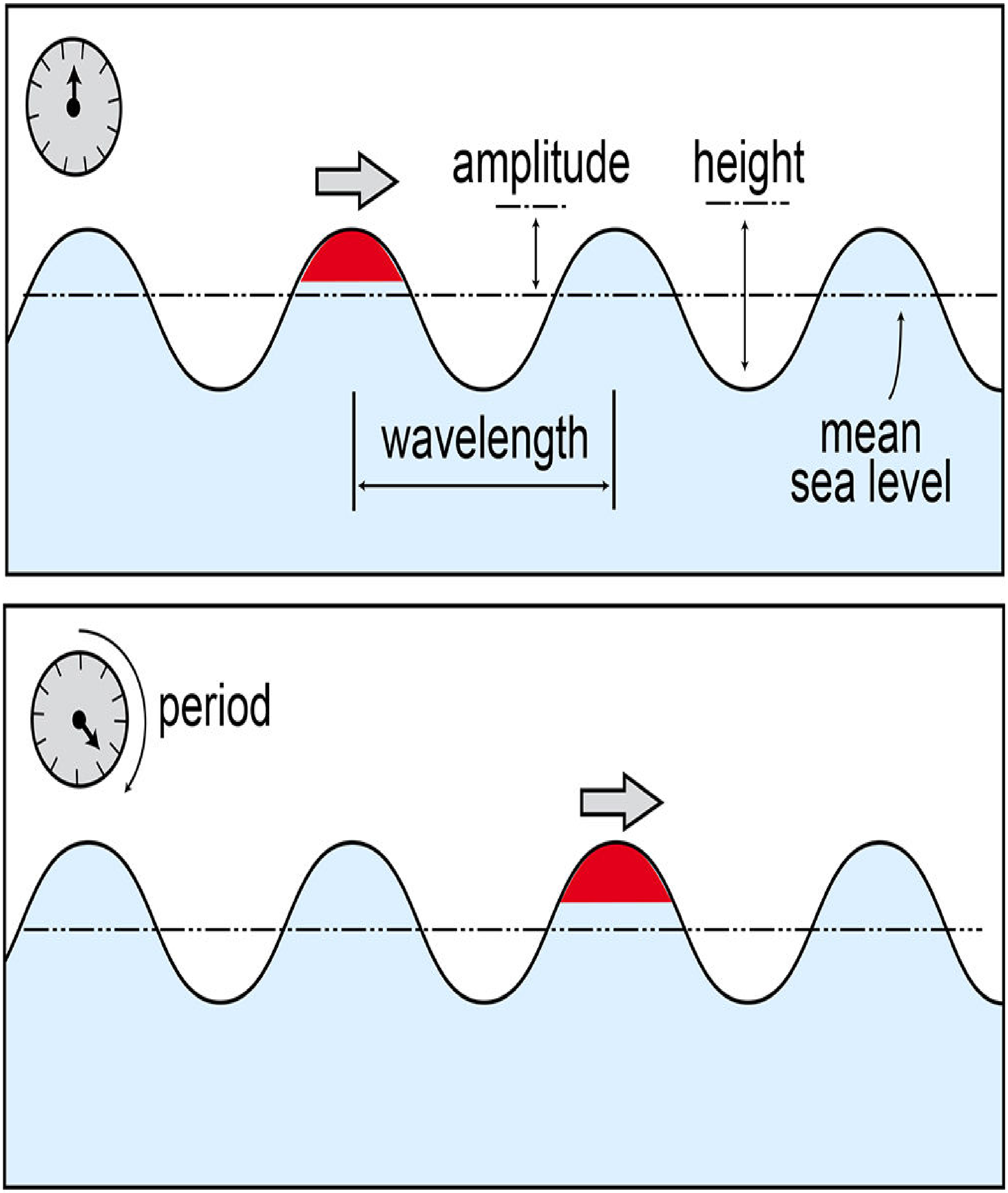

Figure 4.1: Anatomy of a wave.

The Miles-Phillips theory

In the early days, the height of the waves generated by a particular wind blowing across a particular stretch of ocean was calculated **empirically**. This was done by first collecting a large amount of wave height and windspeed data, and then looking for a consistent relationship between one set of data and the other and reducing this to a mathematical formula. Wave heights could then be predicted by ‘plugging in’ a desired windspeed into the formula, without the slightest knowledge of any of the physics behind how the waves were generated by the wind (more on this in Chapter 14).

However, around the mid-1950s, investigative work was underway, in particular by John W. Miles and Owen M. Phillips, to investigate the physical mechanisms behind ocean-wave generation. Miles and Phillips came up with a theory that is still used today in wave-forecasting models. The theory assumes that there are two basic mechanisms involved – the first producing small waves from a completely flat sea; the second then taking over to produce bigger waves from a sea surface already containing small ones. The major difference between these two mechanisms is this: with the first, the energy transfer from air to water is constant with time, so the waves grow at a linear rate; but with the second, the energy transfer rate increases as the waves grow, which makes them grow **exponentially** with time. Eventually, limiting factors intervene to inhibit wave growth, so that a natural maximum wave height is reached for a particular windspeed (Figure 4.2).

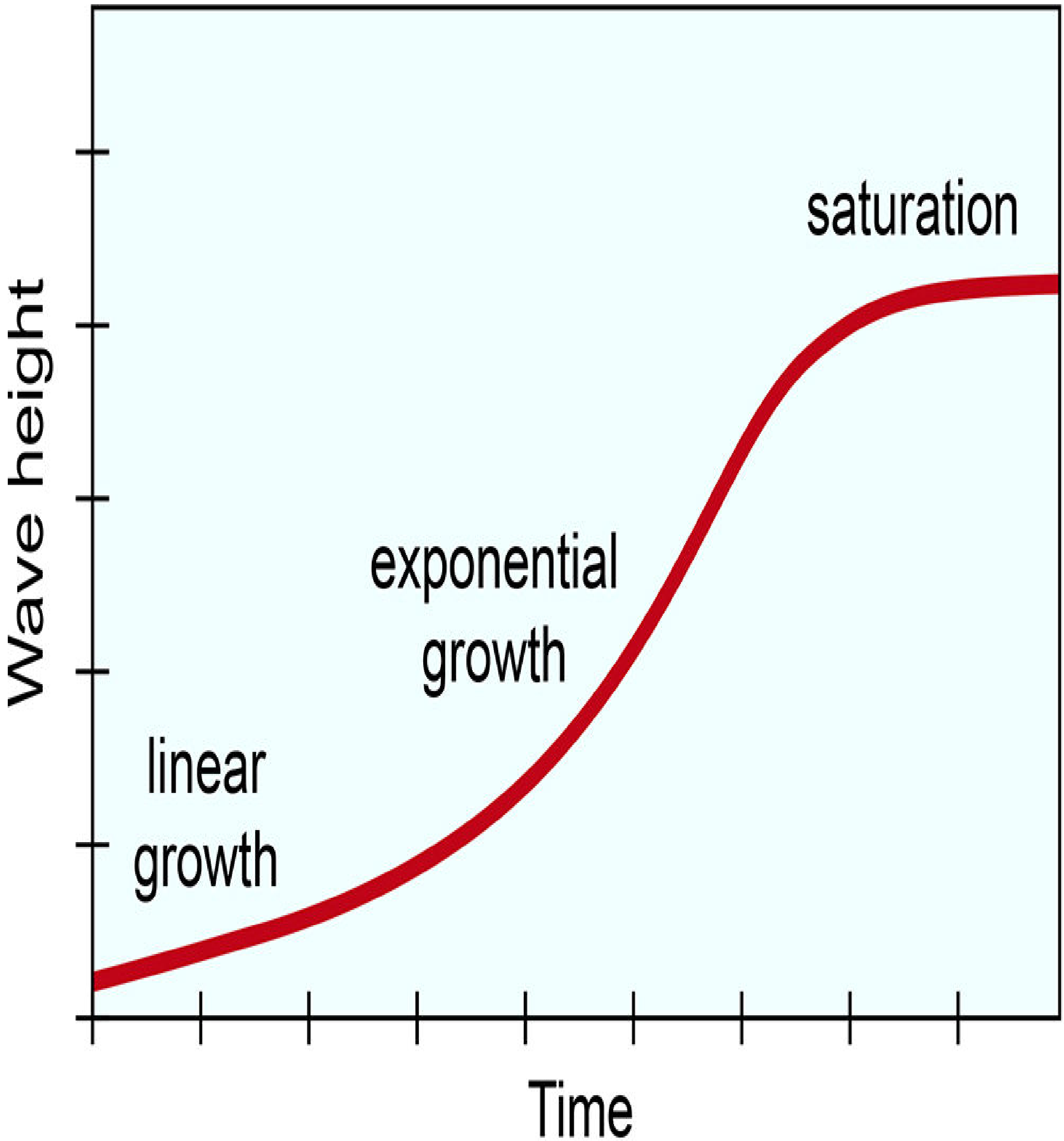

Figure 4.2: Wave height as a function of time in a growing sea.

Linear growth

In 1957, Owen Phillips came up with a theory to explain wave generation from a completely flat sea. The key to this theory is that any wind blowing across the surface of the water never blows perfectly horizontally. The air naturally contains vertical fluctuations as well, causing it to take on a random wave motion above the water surface. As the wind blows across the water surface, the pattern of random ups and downs changes only very slowly, and, therefore, the same pattern is **advected** across the water surface for some distance (Figure 4.3).

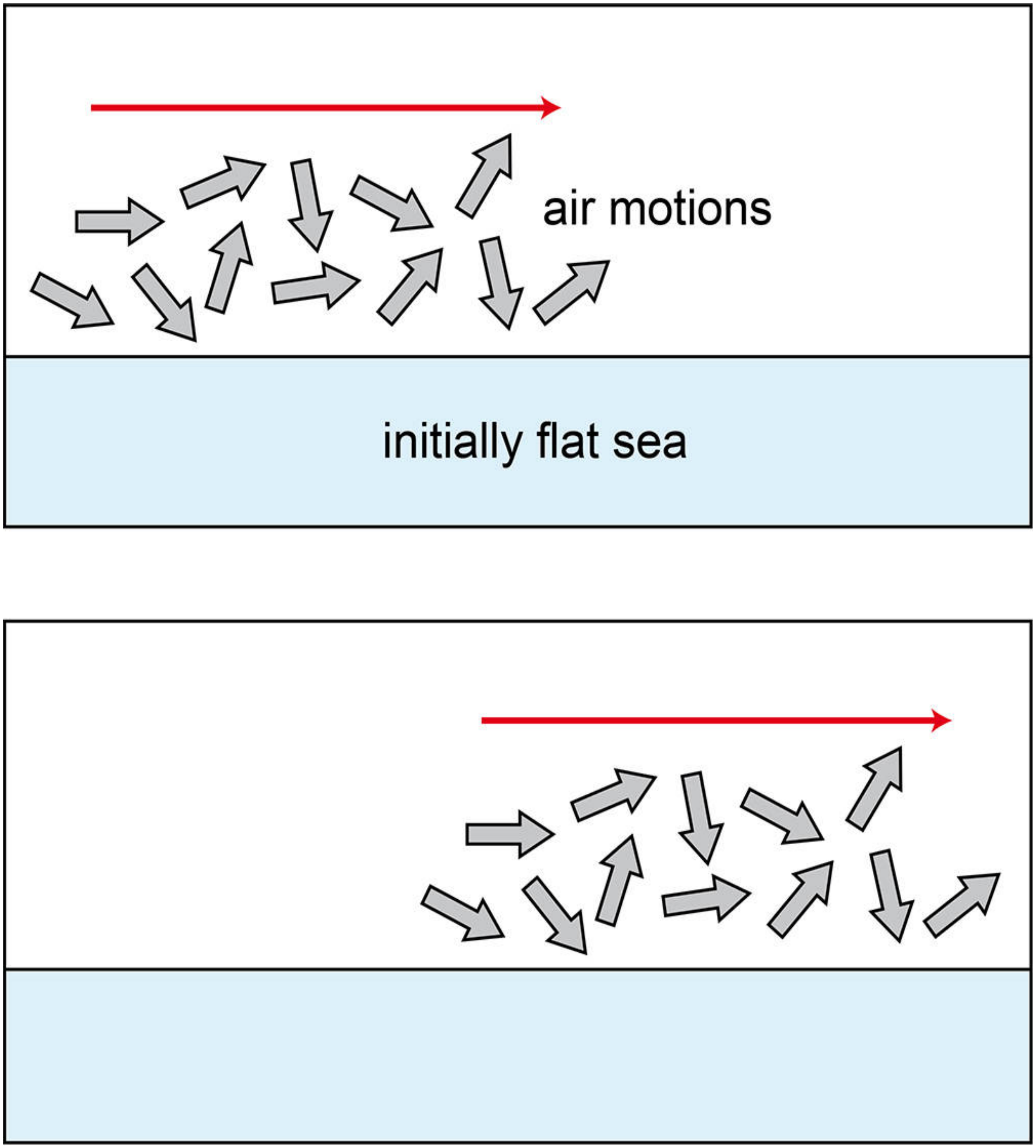

Figure 4.3: The pattern of random air movements shifts across the ocean surface and only changes slowly with time.

As a result, the air pushes down on the water in some places and sucks it up in others, causing the initially-smooth water surface to develop tiny bumps and dips. These bumps and dips move along as the pattern of ups and downs in the air moves across the sea surface. Now, these tiny bumps and dips develop into small waves, which then start to propagate at their own individual speed, independent of the speed of the air disturbance driving them. A certain percentage of these waves, however, remain in sync with the air disturbance. This causes the wave to grow as the air-pressure disturbance follows it and pumps more energy into it (Figure 4.4).

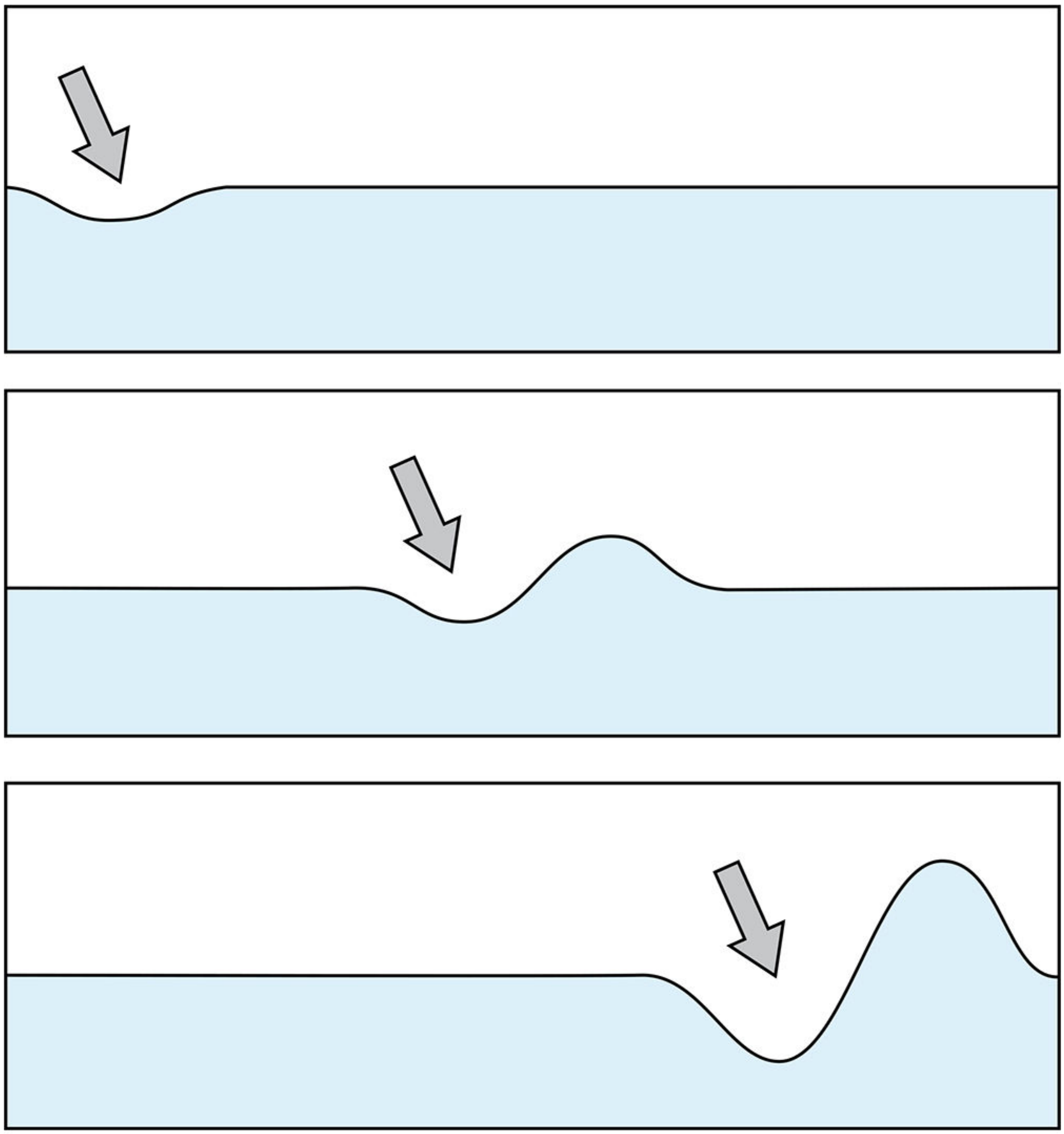

Figure 4.4: If the horizontal speed of a vertical air movement coincides with the wave it produces, the two become synchronized and the wave grows.

However, the majority of waves go out of sync, either immediately or after a short time. As a result, the waves generated by this first mechanism are only able to reach heights and wavelengths of a few centimetres (Figure 4.5).

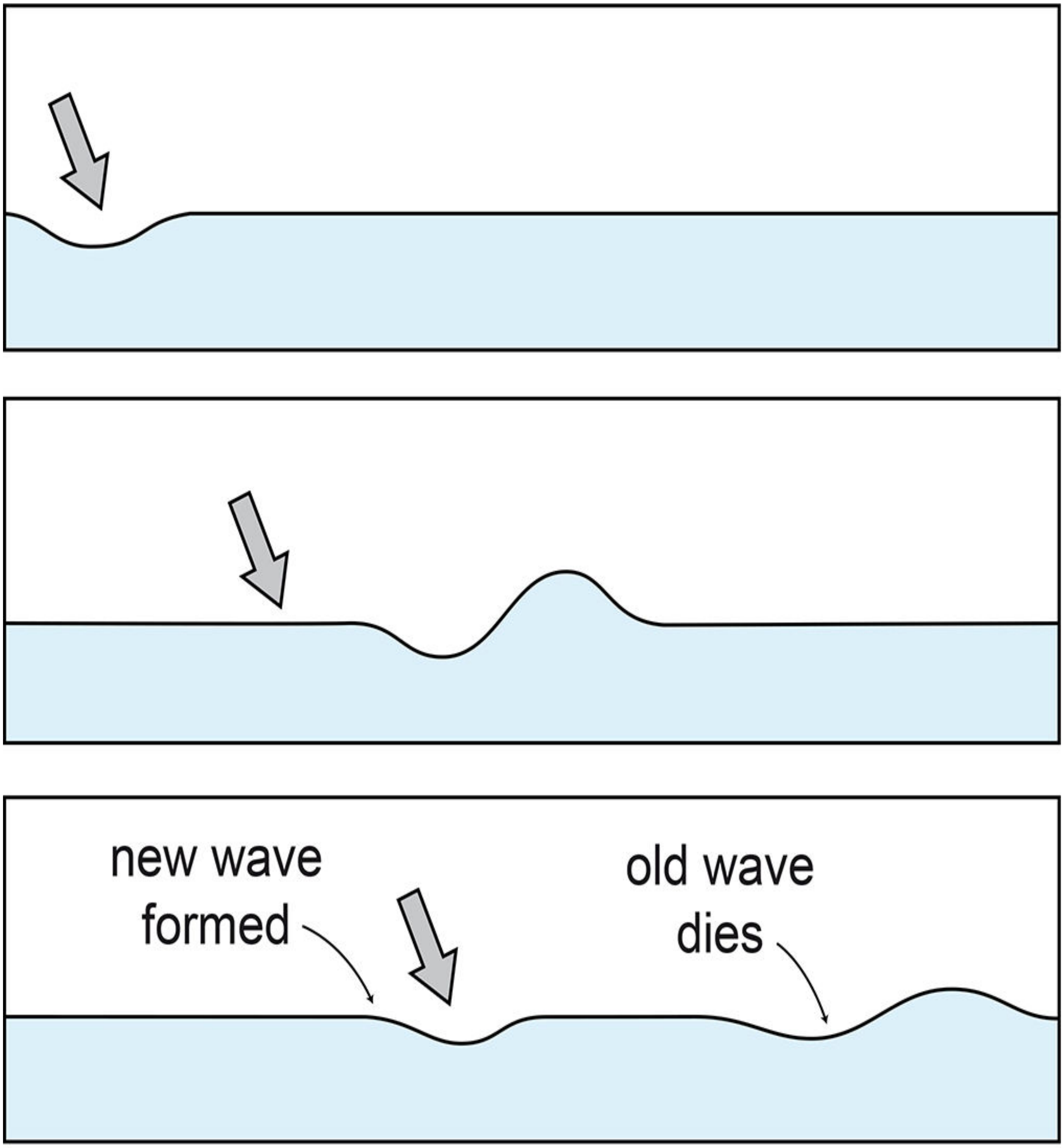

Figure 4.5: Eventually the wave and the air movement that produced it go out of sync, limiting linear wave growth.

Once the sea surface contains these small waves, the second mechanism quickly takes over to produce much bigger waves.

The idea of initial wave growth out of a flat sea is a little academic. The sea surface is never completely flat. There are always *some* waves present, and they are nearly always bigger than the centimetre-high ones produced by the initial growth mechanism. Therefore, most modern wave-forecasting models tend to ignore this mechanism, instead assuming that the sea already contains a 'starting' wave height.

Exponential growth

The exponential growth mechanism was proposed by John W. Miles, also in 1957, extending the work of Phillips and offering a more complete description of how the waves that we see on the ocean surface are formed. The key to the exponential growth mechanism is the fact that as the waves get bigger they begin to interfere with the very air flow that is producing them. As a wave begins to 'stick up' out of the surface of the ocean, it forces the air to ride up over the top of it and back down again. The presence of the wave in the air flow enhances the transfer of energy from air to water. So, the bigger the waves get, the more efficient the energy transfer process becomes, even if the windspeed remains the same. It is a classic **positive-feedback** loop (Figure 4.6).

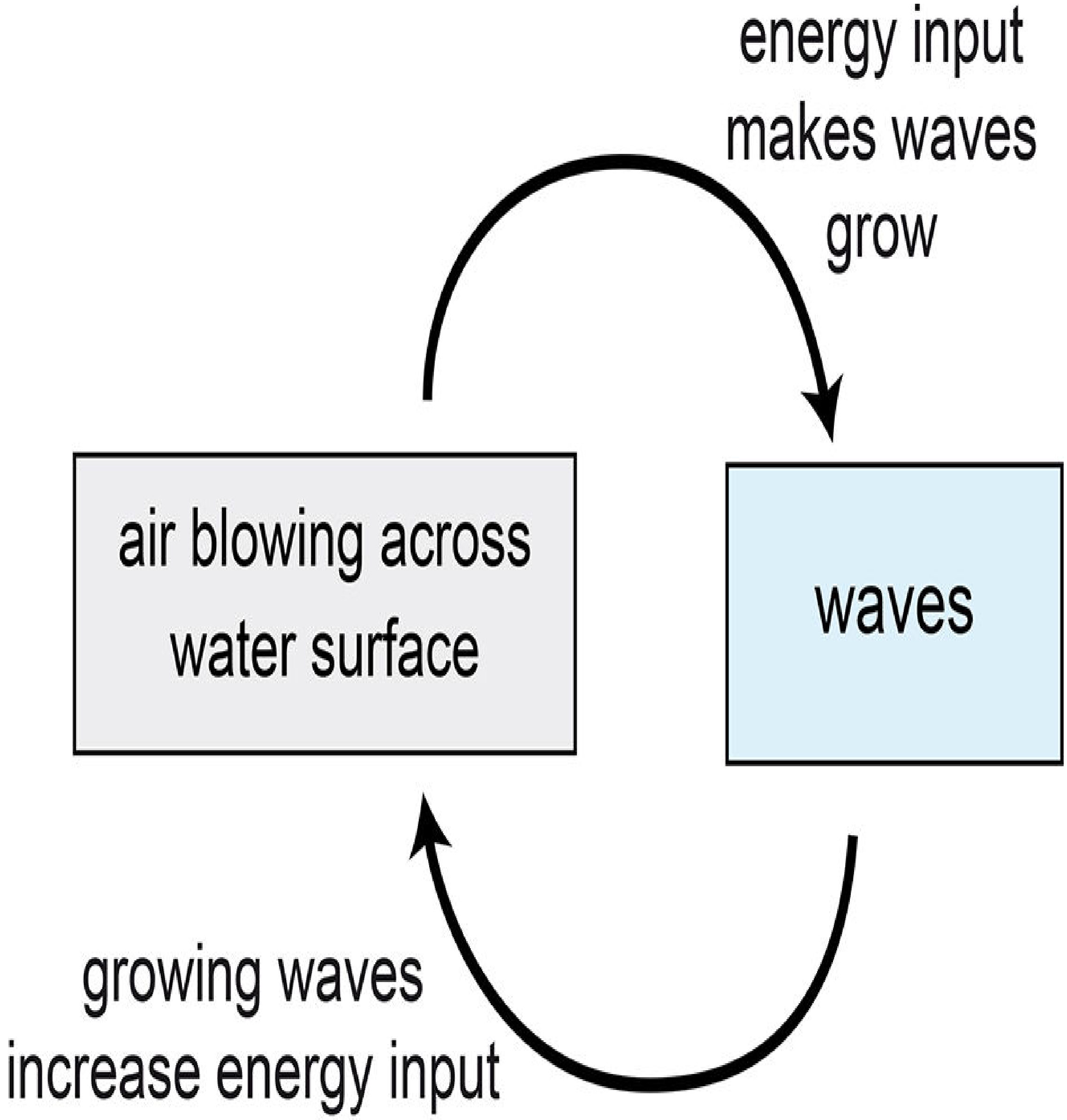

Figure 4.6: The bigger the waves grow the more they influence the air flow, increasing the energy transfer and making them grow faster.

How does this enhancement come about? It is all to do with air pressure differences between the front and back of the wave. Consider a wave travelling along at its own speed, with the wind blowing over the top of it (Figure 4.7, top panel). The air is necessarily travelling faster than the wave itself (otherwise, the wave wouldn't grow). The air pushes against the back of the wave and pulls at the front of the wave. In other words, there is a higher pressure at the back and a lower pressure at the front. This pushing and pulling of the wind transmits energy to the wave and makes it grow. A common way of representing this sort of thing in diagrams is with **streamlines**. In the top panel of Figure 4.7, the streamlines are squashed up at the back of the wave, representing a relatively high pressure, and pulled apart at the front of the wave, representing a relatively low pressure.

As the wave grows and its influence on the windfield becomes greater, the pressure at the back of the wave becomes even higher, and the pressure at the front of the wave becomes even lower. As the wave grows, the streamlines become more squashed up at the back of the wave and more pulled apart at the front (Figure 4.7, middle and lower panels). As a result, the wind pumps energy into the wave at a faster rate, which makes the wave grow even faster, which in turn modifies the windfield even more, and so on. In summary, the bigger they are, the quicker they grow.

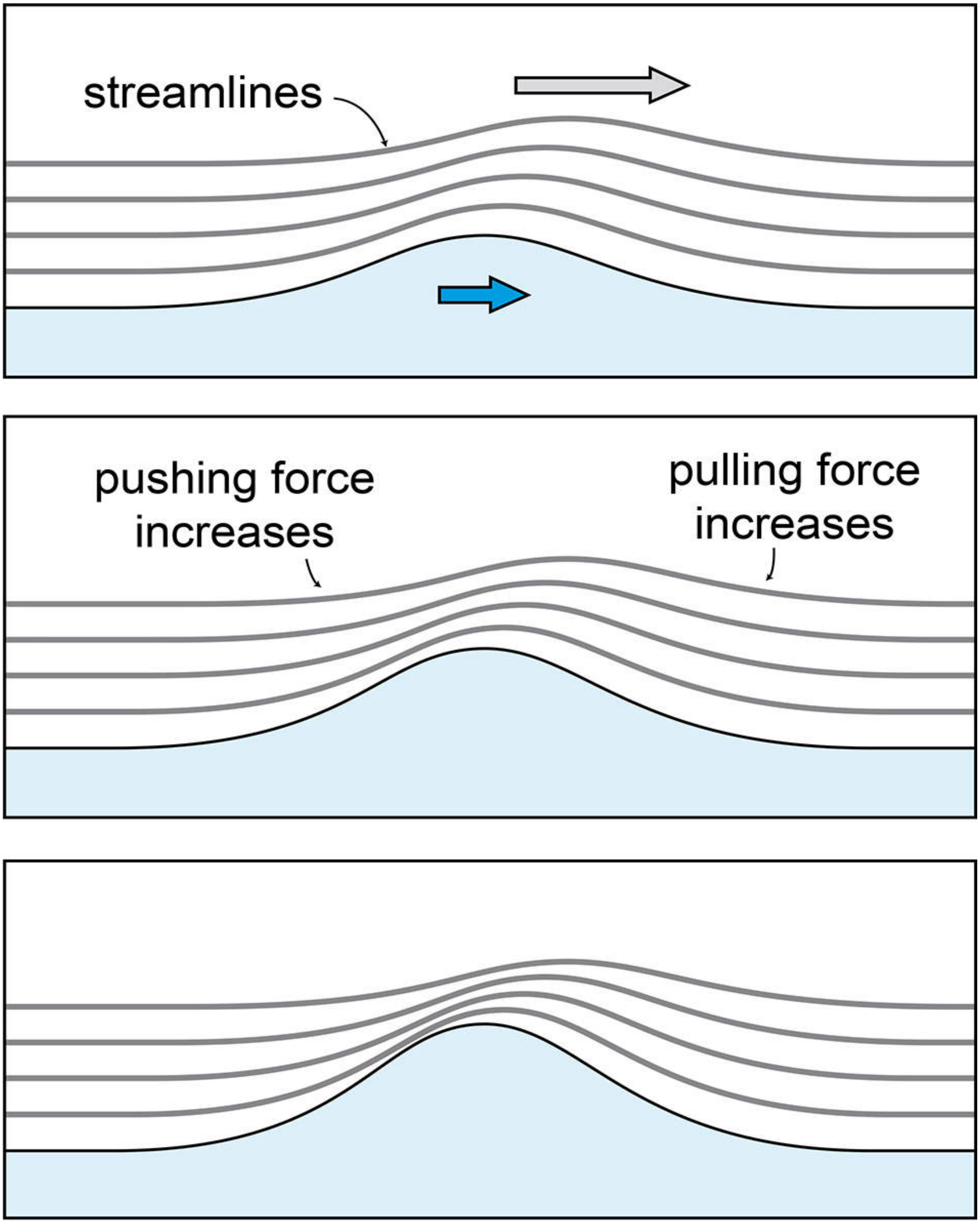

Figure 4.7: Exponential wave growth.

If you don’t believe that the waves themselves have any influence on the air motion, here are a couple of surfing examples. First, think about surfing in a strong offshore wind, particularly if you have ever done so in big waves. If you paddle for a wave and either don’t catch it or decide not to go, for whatever reason, you will feel a tremendous turbulence as the back of the wave goes over. Sometimes the effect is so strong that it literally pushes you off your board. Second, in big waves, in a light offshore or even with windless conditions, it is common to feel an exaggerated offshore wind up the face as you are paddling for the wave. This is because the wave squashes the air in front of it as it moves, accelerating the flow up the face of the wave.

Limiting factors

It would be absurd to think that the waves keep on growing as long as the wind blows. In certain areas of the world, such as in the Southern Ocean, there is a constant strong wind that has been blowing over a large stretch of ocean for thousands of years. According to the theory above, the waves would have already grown to heights of several thousand kilometres or more, which is clearly ridiculous. Therefore, there must be one or more limiting factors that stop the waves growing once they reach a certain height. A balance must be reached whereby the energy input to the waves by the wind equals the energy dissipation by these limiting factors. At this point, the waves reach their limiting height for that particular windspeed and will only grow further if the windspeed increases.

Limiting factors, just like the wave growth processes themselves, are still quite poorly understood. But the one currently considered the most important is **whitecapping**. This is where the waves get so steep in the generation area that the top part momentarily breaks, dissipating a lot of energy in turbulent water motions.

Whitecapping is a familiar sight to almost anyone who has ever seen the sea. Whitecaps are also known by many different names, such as ‘white horses’ in English, ‘little sheep’ in Spanish, or ‘jumping rabbits’ in Japanese. As a surfer you will know that whitecapping always starts happening once the wind reaches a particular strength. In fact, the descriptions of the sea state corresponding to each level in the famous **Beaufort scale** are closely connected to the amount of whitecapping. For example:

> *Force 3:* Large wavelets; crests begin to break; scattered whitecaps
>
> *Force 9:* High waves with dense foam; wave crests start to roll over. Considerable spray
>
> *Force 12:* Huge waves. Sea is completely white with foam and spray…

For each particular windspeed there is a maximum wave height, ultimately reached when the energy sucked out of the waves by mechanisms such as whitecapping becomes equal to the energy put into the waves by the wind. For a given windspeed, once whitecapping begins, the whitecaps themselves get bigger as the waves get bigger. Any further wave growth is quickly brought back down to the limiting value by the whitecaps (Figure 4.8). The limiting of wave growth by whitecapping is a classic **negative feedback** mechanism (Figure 4.9).

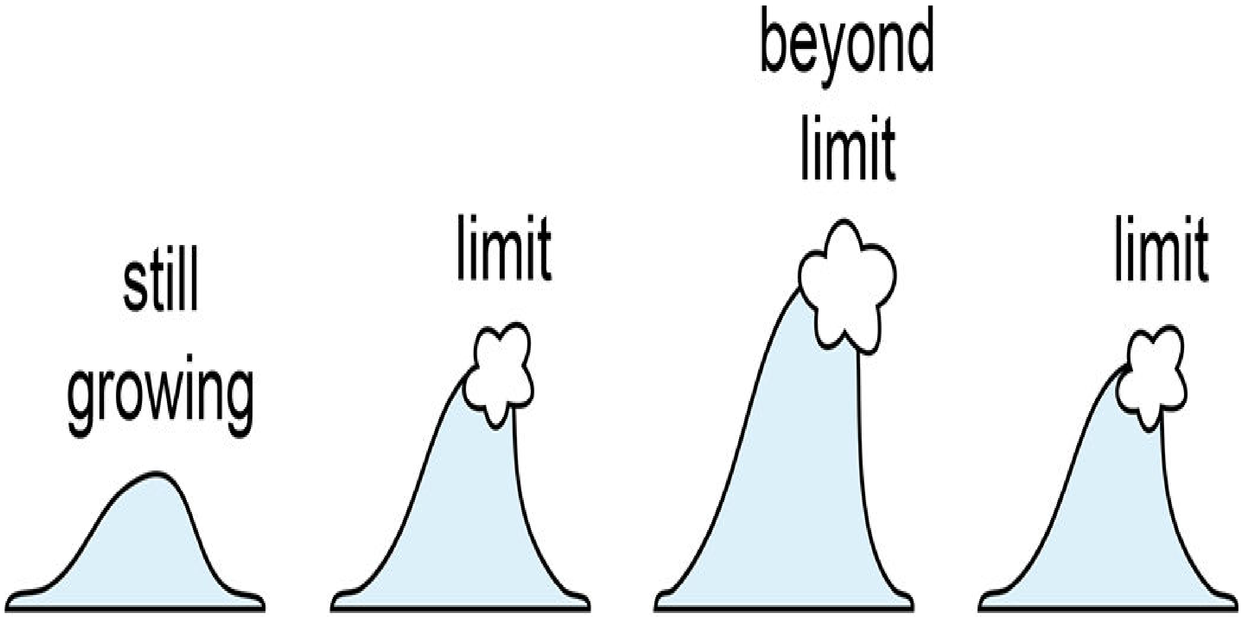

Figure 4.8: Once white-capping starts, the waves cannot grow beyond a certain limit for a given windspeed.

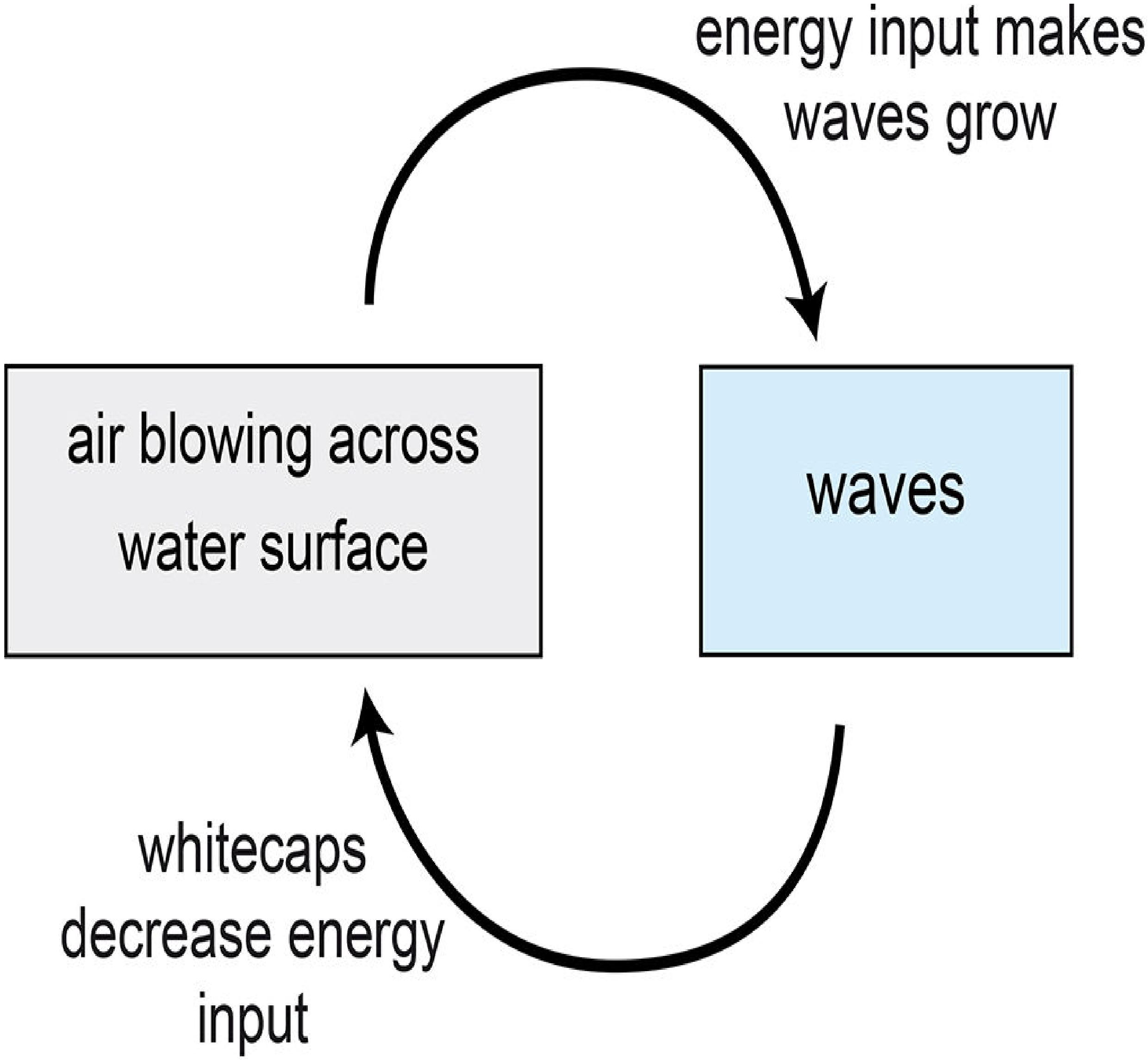

Figure 4.9: The bigger the waves try to grow, the more the whitecapping reduces them, keeping them at a limit determined by the windspeed.

Non-linear transfer

Finally, there is another mechanism that is very important during wave growth, which I will mention here. It is called **non-linear wave-wave interactions**, and relates to the transfer of energy not from the wind to the waves, but between the waves themselves. The reason why it is 'non-linear' is not really important here. What is important is the way in which the different types of waves are affected by this process.

When waves are initially generated by the wind, a mix of different heights, directions and wavelengths are produced. The wavelength is the distance between one wave crest and the next. (Strictly we should use the word **period** – the time taken for one complete wave to pass a fixed point – but for a simple explanation, it doesn't matter.)

That is why, in a storm, the sea surface looks like a confused mess. It is between the waves of different wavelengths that the process of non-linear transfer takes place. It robs energy from waves of some wavelengths and gives it away to waves of other wavelengths. There is no net loss or gain of energy in the sea due to this process.

The sea contains a smooth continuum of wavelengths, with an infinite number of different wavelengths between the shortest and the longest. Here, for simplicity, we will split that continuum into just three categories – short, medium and long waves. Due to the non-linear transfer process there is a large, continuous flux of energy from the medium-length waves to the long waves, with a small amount also being transferred from the medium to the short waves. The reason why this happens is very complex, but we can think of it simply as the long waves gobbling up the medium ones as they grow.

Most of the waves generated by the wind have short or medium wavelengths, not long ones. That is to say, the energy input from the wind principally affects the short and medium waves. Likewise, the energy dissipation due to whitecapping principally affects the short waves (because the short ones 'stick up' out of the sea surface most). The long waves suffer very little energy loss due to dissipation. Therefore, taking all these energy inputs and outputs into account, we can see that the long waves receive a greater net amount of energy than any of the others (Figure 4.10).

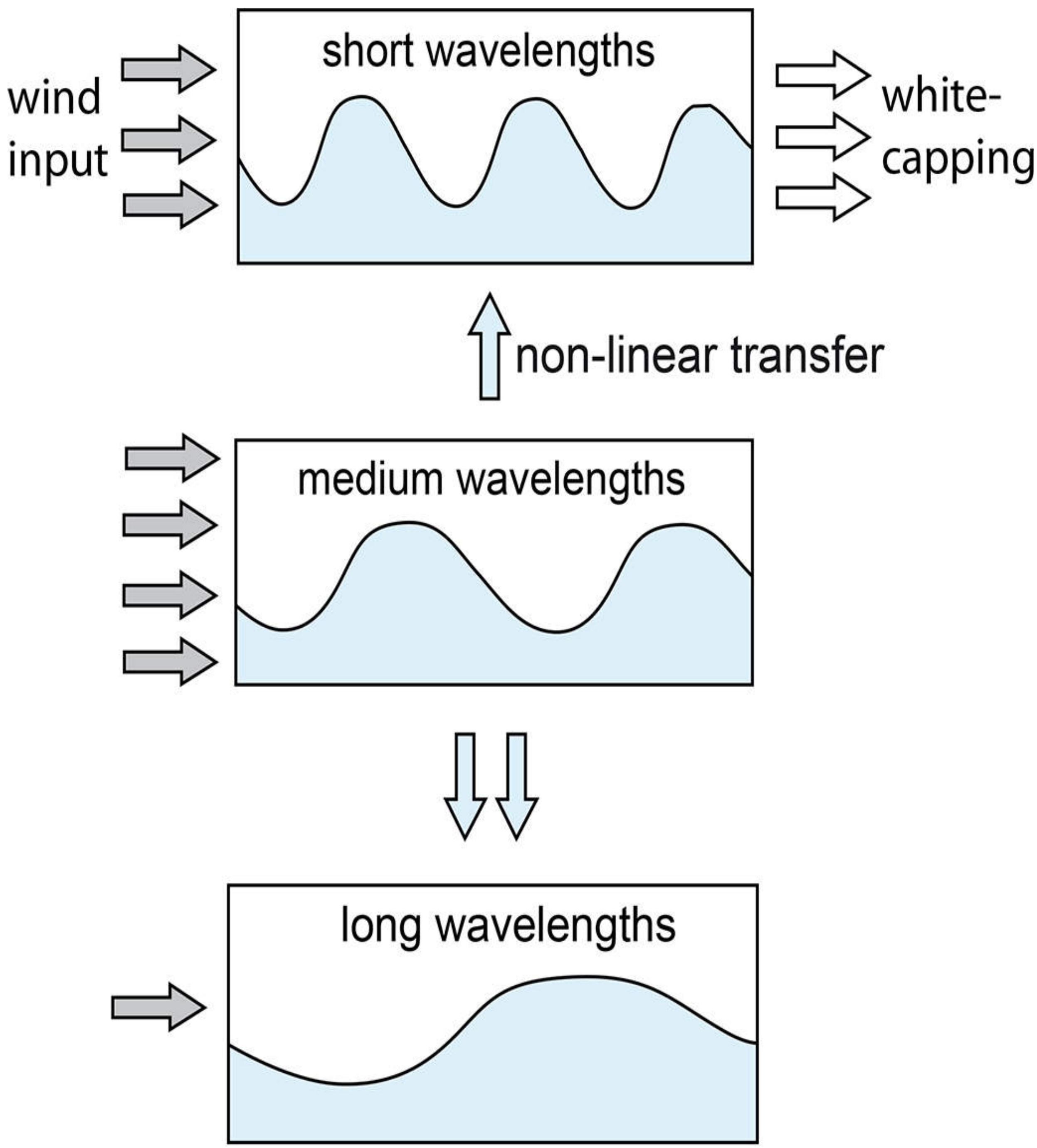

Figure 4.10: Energy transfer in a growing sea. The number of arrows represents the amount of energy transferred.

As the sea state grows in the storm, the sea becomes progressively more dominated by the long waves. That is why, in a growing sea, the average wave height increases, but the average wavelength also increases slightly at the same time.

5 Propagation of Free-travelling Swell

Introduction

So far, we have looked at how waves are produced by the wind blowing across the surface of the ocean – wind that is the result of air trying to rush from a region of high pressure into a region of low pressure. The area over which the wind is blowing – the wave-generating area – is more or less the location of the low pressure itself, and is referred to as the **storm centre**. Here the waves are constantly supplied with energy by the moving air, and the sea contains a wide range of waves of all different sizes, lengths, shapes and directions. This is known as a **windsea**.

However, once the waves start leaving the generating area, they no longer remain under the influence of the overlying wind, and propagate away as free-travelling **swell**. In this chapter we look at the properties of free-travelling swell and the ways in which the waves change as they propagate further and further away from the storm centre (Figure 5.1).

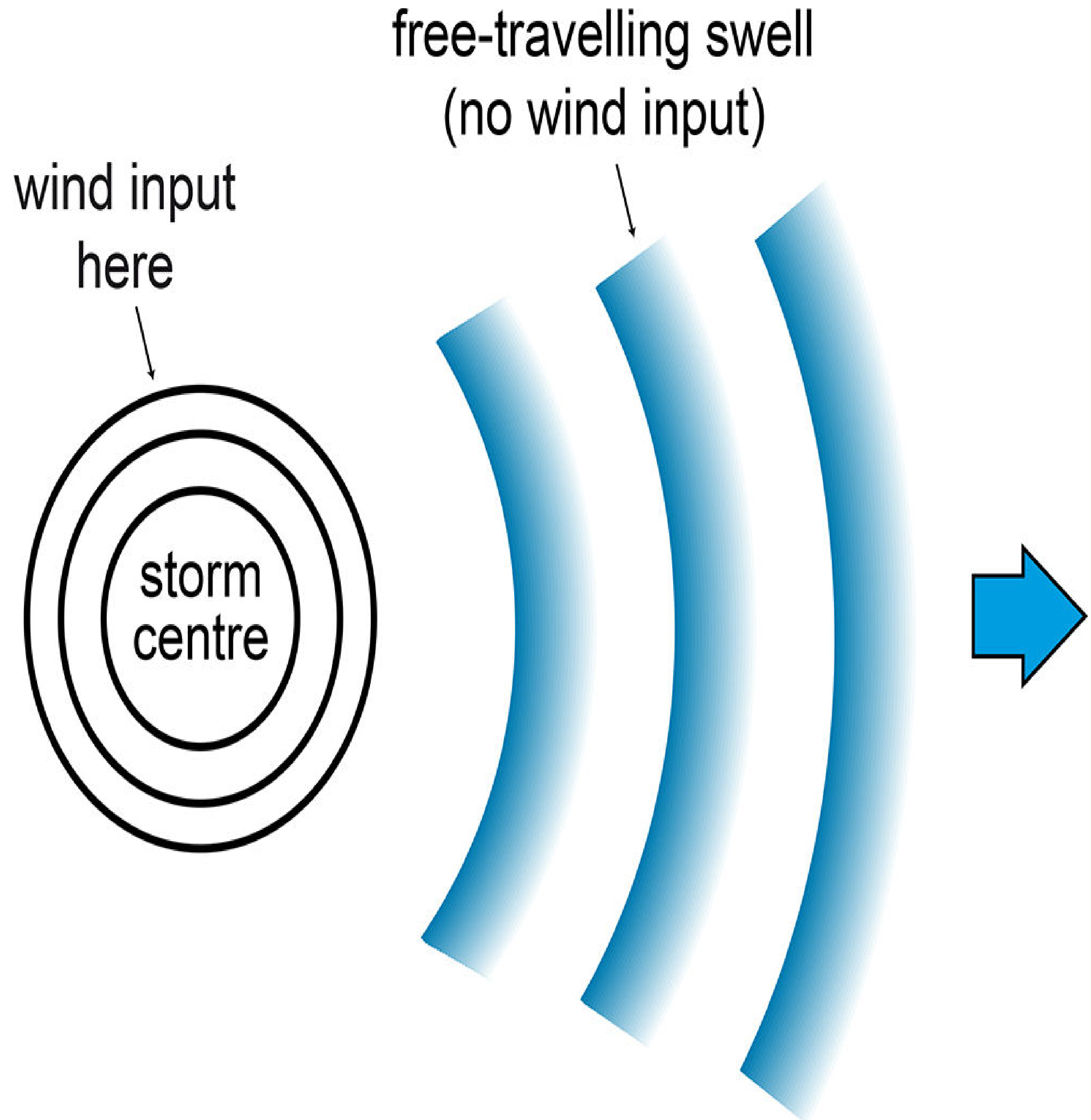

Figure 5.1: The wind transfers energy to the waves in the storm centre only. Outside the storm centre the waves are free-travelling.

If the storm centre is some distance away, the waves we ride on the beach will be quite different from those originally generated. Even before the waves start to hit the shallower water of the continental shelf, several things happen to them as they propagate across the ocean. As the swell moves away from the generating area it not only spreads out over a progressively wider area – **circumferential dispersion** – but also stretches out in the direction of propagation – **radial dispersion** (Figure 5.2).

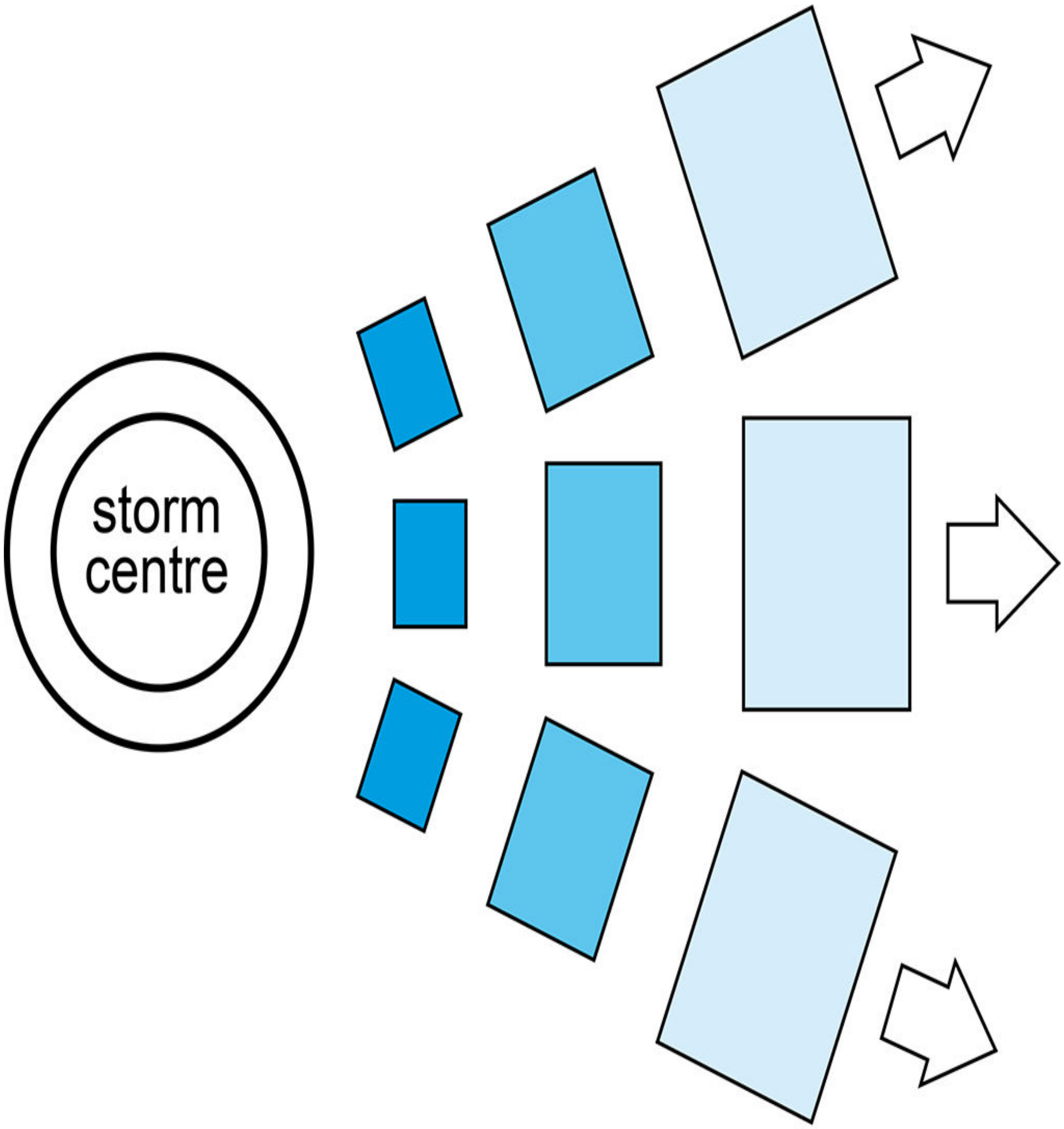

Figure 5.2: Swell propagating away from the storm centre. The further away it gets, the more it expands radially and circumferentially.

Also, throughout the propagation path, the waves sort themselves out into groups or sets. The particular characteristics of these wave groups as they arrive on our reefs and beaches can make a great difference to our surfing experience on any given day.

Very little energy is actually lost when oceanic swell travels on its own and, although the waves do get smaller through circumferential dispersion, they can be detected many thousands of kilometres away from the storm centre. In theory, swell could probably travel all the way around the world if the continents weren't there to block it.

Swell: a messenger of energy

When we think of waves travelling over the surface of the ocean, we must remember that, in deep water anyway, the waves do not 'carry' water from one place to the other. They are not like ocean currents, where water from the Antarctic can end up in the Atlantic. Waves are simply messengers of energy.

This is easily demonstrated by taking a piece of carpet and flicking it up and down so that a wave propagates from one end to the other. The energy needed for the up-and-down movement is transmitted along the carpet, but the carpet itself doesn't go anywhere. The wave does not 'carry carpet' from one end to the other; it just transmits energy. Each particle in the carpet describes a circular motion as the wave passes through it, eventually ending up in the same place. The wave is effectively carrying a message through the carpet, telling the particles to go round in a circle. The fact that waves can be used to carry a message is the fundamental principle behind sound, radio and almost every kind of communication method known.

In the deep ocean, the sea surface behaves just like a carpet. Any floating object, if you looked at it from the side, would trace a complete orbit every time a wave passed through, ending up in exactly the same spot after the wave had passed. The wave transmits a message from one part of the ocean to the other, telling the particles on the surface to go round in a circle (Figure 5.3).

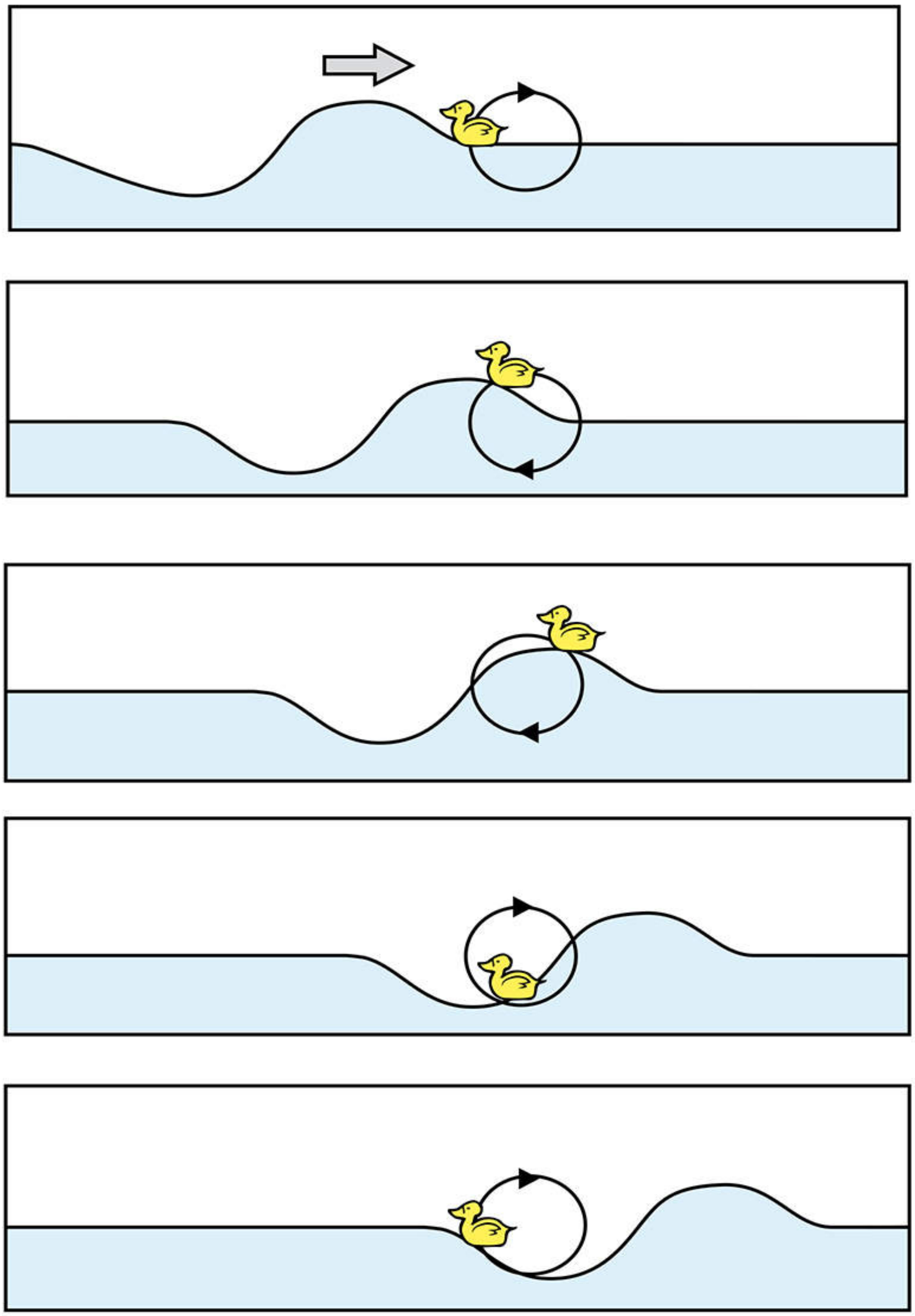

Figure 5.3: In deep water, as a wave passes a floating object, the object traces a circular path and ends up in the same place.

The motion of the particles on the sea surface is also transferred to the water below. Because of friction and energy loss between vertical levels in the sea, the orbits of the particles within the water get smaller with distance from the surface. Out in deep water, these orbital motions diminish to nothing before they reach the sea-bed (Figure 5.4).

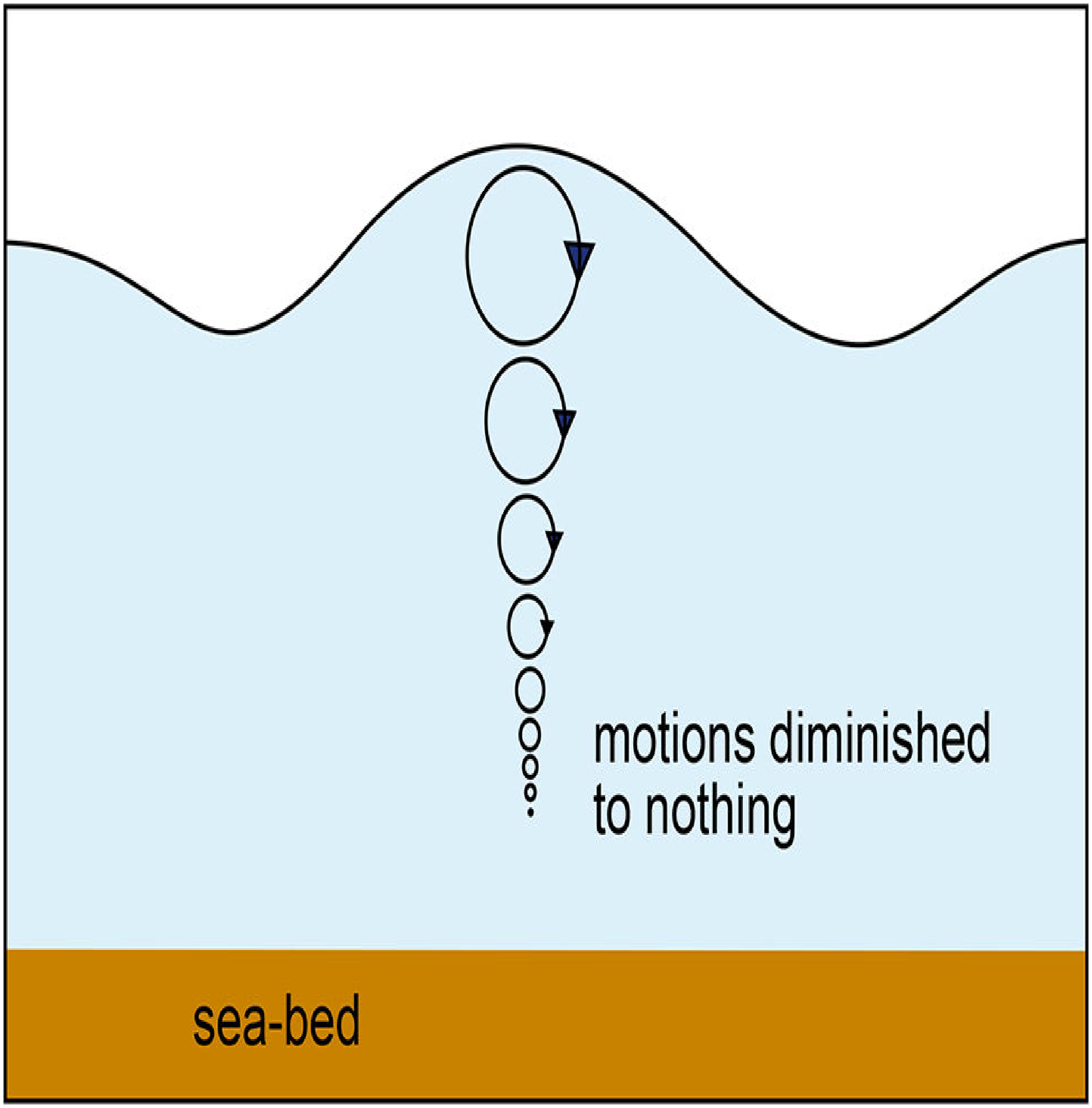

Figure 5.4: In deep water, the orbital motions beneath a wave diminish to zero before reaching the sea-bed.

Another important property of swell is that it is not affected by the Coriolis force (the Coriolis force is what makes moving objects turn to the right in the Northern hemisphere and to the left in the Southern hemisphere – see Chapter 2). Swell does not ‘bend’ as it travels across the surface of the planet, because, unlike ocean currents, it does not carry any physical material from one place to another, just a message.

Circumferential dispersion

As the swell travels away from the storm centre, it spreads out over a progressively wider area. As the swell propagates over the ocean surface, the movement of the water molecules is transmitted to other neighbouring molecules. Therefore, the wave-fronts become ever wider, spreading the same amount of energy over a greater area. The spreading of the energy ‘robs’ height from the waves, so unless more energy is pumped into the waves to compensate, the wave height will get progressively smaller with distance from the storm centre.

To find out how much smaller the waves get as they spread out, we can use the following principle. We will assume that the waves originate from a single point source on the ocean. The width of the wave-front is approximately proportional to the distance it has travelled away from the point source. For instance, when the swell is 2,000 km from the storm centre, it will have spread out twice as far as when it was only 1,000 km away. So the wave energy 2,000 km away from the storm centre will be spread out over twice the width, therefore it will be half as concentrated as it was 1,000 km from the storm centre. Owing to the relationship between wave energy and wave height, an energy reduction of half means a height reduction of about 30 per cent. Therefore, for every doubling of the distance from the storm centre, the wave height is reduced by about 30 per cent (Figure 5.5).

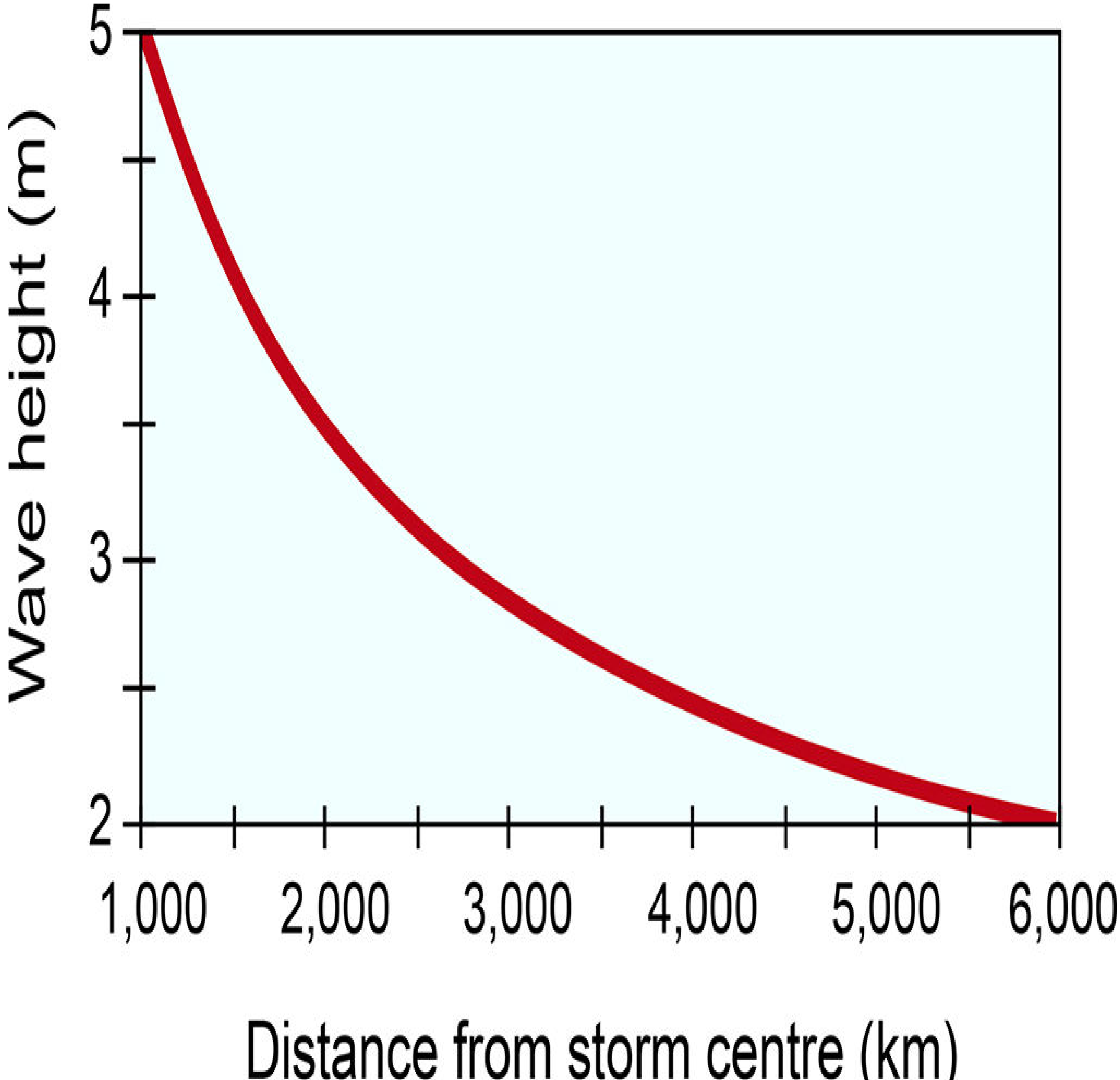

Figure 5.5: Circumferential dispersion makes the wave height diminish with distance from the storm centre.

As a theoretical example, think of a low-pressure system centred just south of Iceland, pumping swell towards Europe. Again, we will assume that the low pressure is a point source, with the swell radiating out from a single point on the ocean. As the swell leaves this point, it starts to spread out. By the time it reaches a spot 1,000 km away – somewhere in Ireland perhaps – the waves arriving on the coast might be, say, 5 metres high. Some of the swell ends its life here, the waves giving up their energy on the reefs, points and beaches of western Ireland. But a great deal misses Ireland altogether, and continues on its journey southward. When it reaches a point 2,000 km from the storm centre – somewhere along the north coast of Spain, perhaps – the wave height will be down to 30 per cent less than it was in Ireland – to about 3.5 m. When it has travelled another 2,000 km, to the Canary Islands, for example, the wave height will be down to 30 per cent less than it was in Spain – to about 2.3 m. The swell would continue losing height in this way as it got further and further away from the point source, getting ever smaller but never quite reaching zero (Figure 5.6).

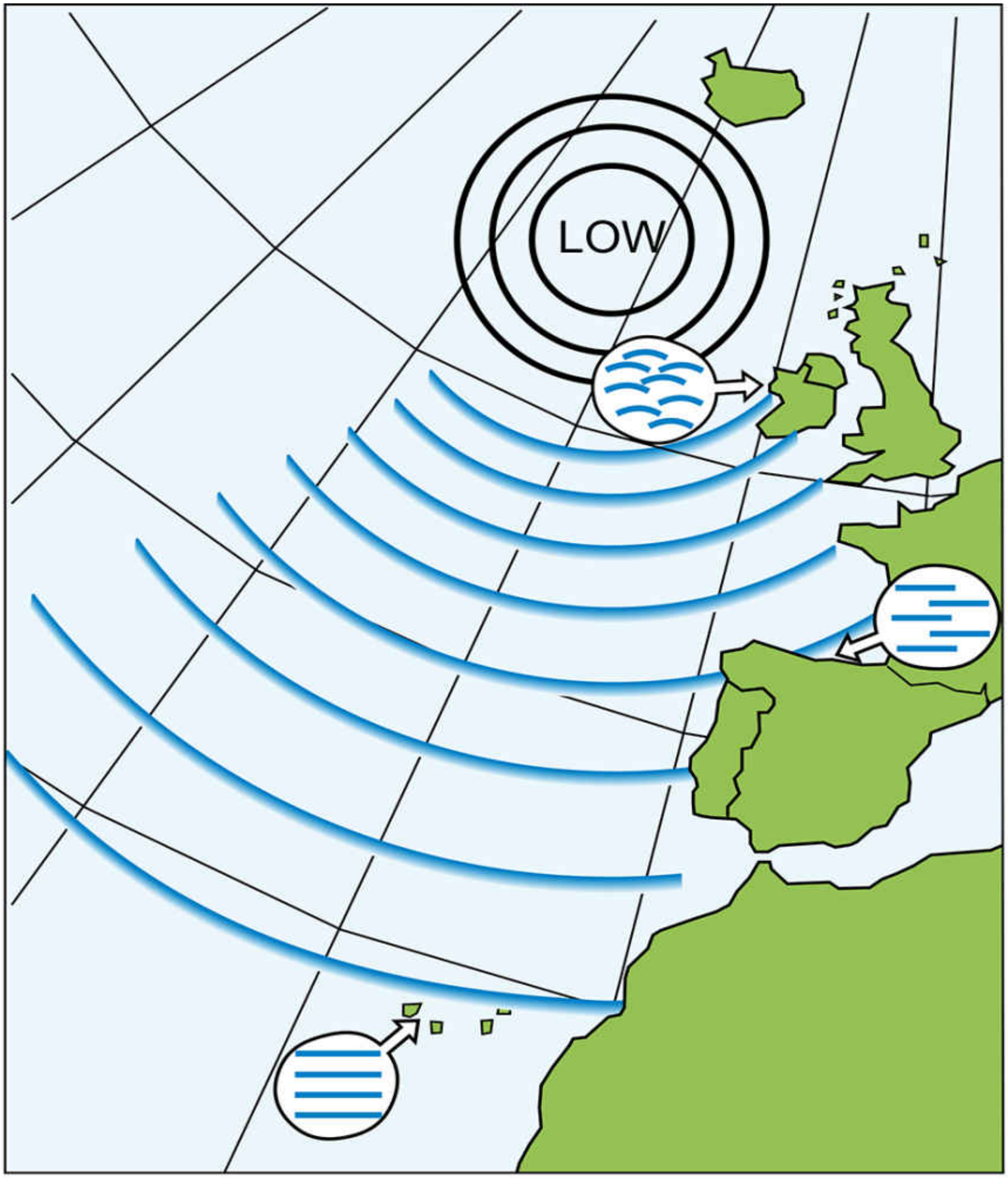

Figure 5.6: A swell generated near Iceland progressively loses height but cleans up as it reaches Ireland, Spain and the Canaries.

Of course, unlike our example, real storms are not singular point sources of swell. In reality, the wind in the storm blows over an area of ocean – the **fetch** – that can be quite wide. As a result, the spreading out of the wave fronts has less effect on the energy reduction with distance. The overall reduction in wave height for every doubling of distance from the storm centre could be as little as 15 to 20 per cent, depending on the width of the fetch. So, if the fetch is particularly wide, the waves that arrived in Spain and the Canary Islands would probably be bigger than in our example. In the Pacific and Indian oceans, storms tend to cover even wider areas than in the Atlantic, so the corresponding wave height reduction in those oceans could be even less.

Also, as the waves get further away from the storm centre they 'clean up', so the waves hitting Spain would be cleaner than those hitting Ireland, and the ones hitting the Canary Islands would be cleaner still. I will explain this below.

Radial dispersion

One of the things that makes water waves so complicated is that they do not all travel at the same speed. Their speed is governed by their period – the time taken for two successive wave crests to pass the same point. The longer the period, the faster they go.

> In deep water, the speed in metres per second at which the swell itself travels across the ocean is approximately 0.78 times the wave period in seconds.

This is **radial dispersion**, probably the most fundamental principle of the propagation of surface waves across the deep ocean. It tells us that waves of different periods **disperse** when they travel away from the storm centre. They all start off together, but then the longer ones begin to outrun the shorter ones until they are all clearly separated out along the propagation path.

A good analogy for this might be a group of marathon runners, some of them having longer legs than others, and therefore being able to run that much faster. Having all started off together, the longer-legged ones eventually pull out in front, leaving the shorter-legged ones behind.

When a swell is first generated, in the storm centre, all sorts of waves are produced at the same time. Very close to the storm centre there is a mixture of many waves of different sizes, shapes, wavelengths and directions. As the swell begins to propagate away from the storm centre, the different waves begin to sort themselves out, the longer, faster ones racing out in front, and the shorter, slower ones lagging behind. The swell has 'stretched out' in the direction of propagation – radially.

Applying this to the same practical example above, if you were very close to the storm centre (Ireland, for example), you would see almost the entire swell including all the long and short waves arriving in a very short time. In contrast, if you were, say, 2,000 km away (Spain, for example), the swell would have had a chance to disperse considerably. So the longest waves, now being much further out in front, would arrive before all the others. After the longest ones, all the other waves would turn up, with the shortest ones eventually coming in last. The swell in this case would be cleaner and more lined up, with a smaller number of different periods arriving at the same time. Even further away (say 4,000 km at the Canary Islands) the swell would be even more dispersed and the different periods would be even more separated out from each other (Figure 5.7).

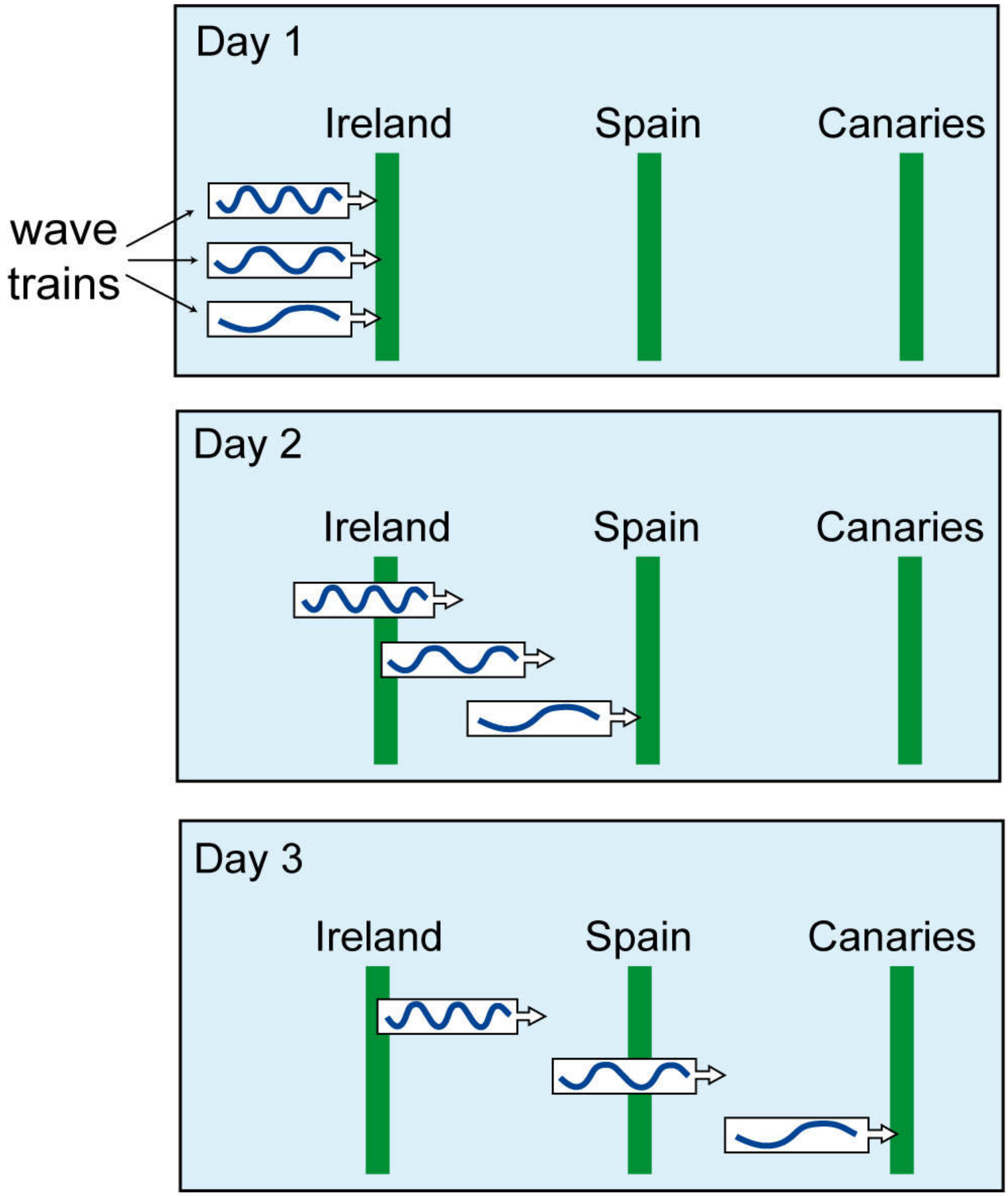

Figure 5.7: Radial dispersion with three wave-trains of different periods.

If you are some distance from the storm centre, the first waves of a new swell are the longest ones. However, the first ones are not always the biggest – sometimes the peak in wave height arrives several hours after the first waves of the swell. The arrival of a new, long-travelled swell is always very interesting to watch. Long lines start appearing, usually very straight and regular, barely detectable at first, but with a noticeable separation between one wave and the next. Within a couple of hours, the waves could be huge.

After the swell has peaked in size, the waves do not have the same punch to them as they did before it peaked. Right at the end of the swell, when the very last waves are coming in, they are normally relatively powerless. Sometimes the shortest waves never even make it to the coast. Short waves are steeper than long ones, so they 'stick up' more out of the ocean surface, and are more susceptible to having their energy removed by things like opposing winds in the propagation path, or by *whitecapping* (Chapter 4). Most of the attenuation of short waves tends to occur near the storm centre.

The formation of groups

Somewhere in the propagation path, between the storm centre and the coast, the waves organize themselves into sets – groups of two, three or maybe ten waves, with much smaller waves, or nothing, in between. The particular manner in which waves are organized into sets is extremely important if you are a surfer. The quality of a surf session can depend to a great extent on the number of waves per set, how far apart the sets are, or which wave in the set is the biggest.

Normally, the further you are from the storm centre – and, hence, the more dispersed the swell is – the more opportunity the group-forming mechanism has to work on the waves. Very close to the storm centre, the sets are hardly noticeable – even though there are probably bigger and smaller waves around, they arrive more randomly and continuously. But if you are thousands of kilometres away, there will be long stretches of time where the ocean appears completely flat, interspersed with the arrival of very clean and lined-up sets of regular numbers of waves.

The fundamental mechanism that makes the waves form into sets is actually quite simple. It stems from the interference between waves of different wavelengths, and is called **linear superposition**. In reality, many different waves of different sizes, lengths and directions are produced in the storm centre, and the interaction between all these along the propagation path is highly complicated. If you include the effects of radial dispersion where waves of different wavelengths travel through the deep ocean at different speeds, things start to get nightmarish.

To keep things simple, we will talk about what happens when waves of just two different wavelengths interact with each other, both travelling in the same direction. It is a highly simplified case, and quite unrealistic, but nevertheless it is still a valid way to understand the basic mechanism responsible for wave groups.

Imagine a wave train containing waves of a certain wavelength travelling in a certain direction. You then take another wave train containing waves of a slightly different wavelength, travelling in the same direction, and superimpose them on to the first one. The two wave trains interfere with each other, and a resultant wave train emerges, being the sum total of the two original ones. Since they are of different wavelengths, they do not just add together at every point. Where the peaks or troughs coincide, you get a bigger peak or trough – called **constructive interference** – and where a peak and a trough coincide, they cancel each other out – called **destructive interference**. The resultant wave train then looks like a primitive wave group (Figure 5.8).

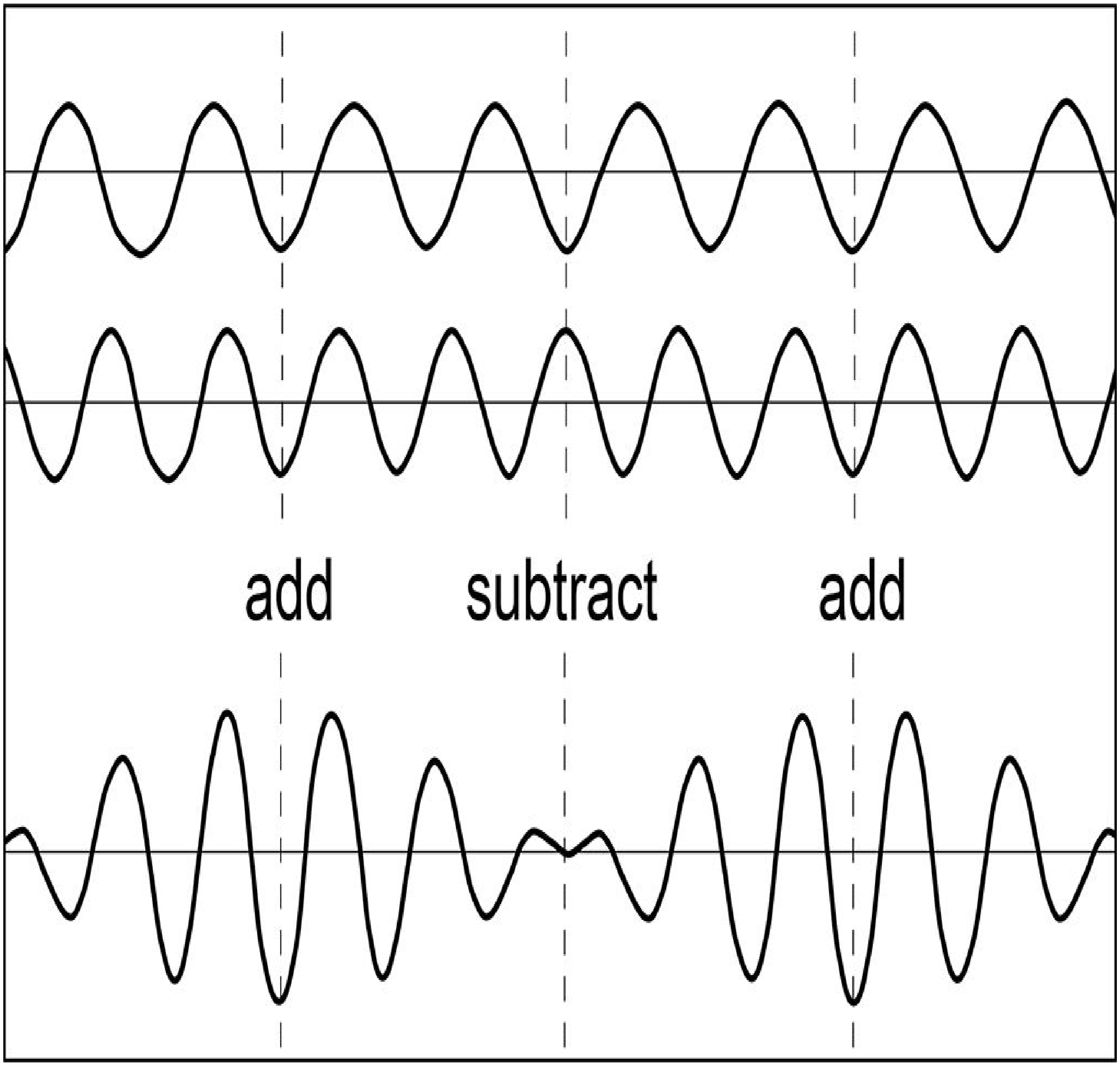

Figure 5.8: Snapshot of two superimposed wave trains of different wavelengths. The result (bottom) is a primitive wave group.

There is something very curious here that I'll mention briefly. The simple example I have given is a 'snapshot' of the ocean surface, 'frozen' in time. So, Figure 5.8 could be thought of as a graph of ocean surface elevation as a function of horizontal distance along the water surface. This is not the same as a plot of ocean surface elevation as a function of time, at a single point on the ocean surface, for example if you plotted the up-and-down motion as a function of time of a floating object moored in the same spot, as a wave train passed over it. Now, with exactly the same grouping structure, the number of waves in a group is different according to whether you looked at them as a function of time or distance. The reason is to do with radial dispersion and the fact that waves of different wavelengths travel at different velocities. I'm not going to say any more about that, as it is beyond the scope of this book.

Wave speed and group speed

The speed of each individual wave in deep water is twice the speed of the group of waves that it is travelling in. That sounds a bit weird if you haven't heard it before, so let me explain. The group moves across the ocean at a speed called **group speed**, but each individual wave in the group travels at twice the speed of the group, at a speed known as **phase speed**. As the group is crawling along the surface of the ocean, all the waves in it are constantly moving from the back of the group to the front. This is a bit like taking a moving staircase, laying it flat on the ground while it is still going, and then pulling the whole thing along the ground. Each stair would appear at the back, travel to the front of the staircase, and then disappear. This is what happens with waves in a group. Waves are born at the back of the group, move through the group to the front, and then die (Figure 5.9).

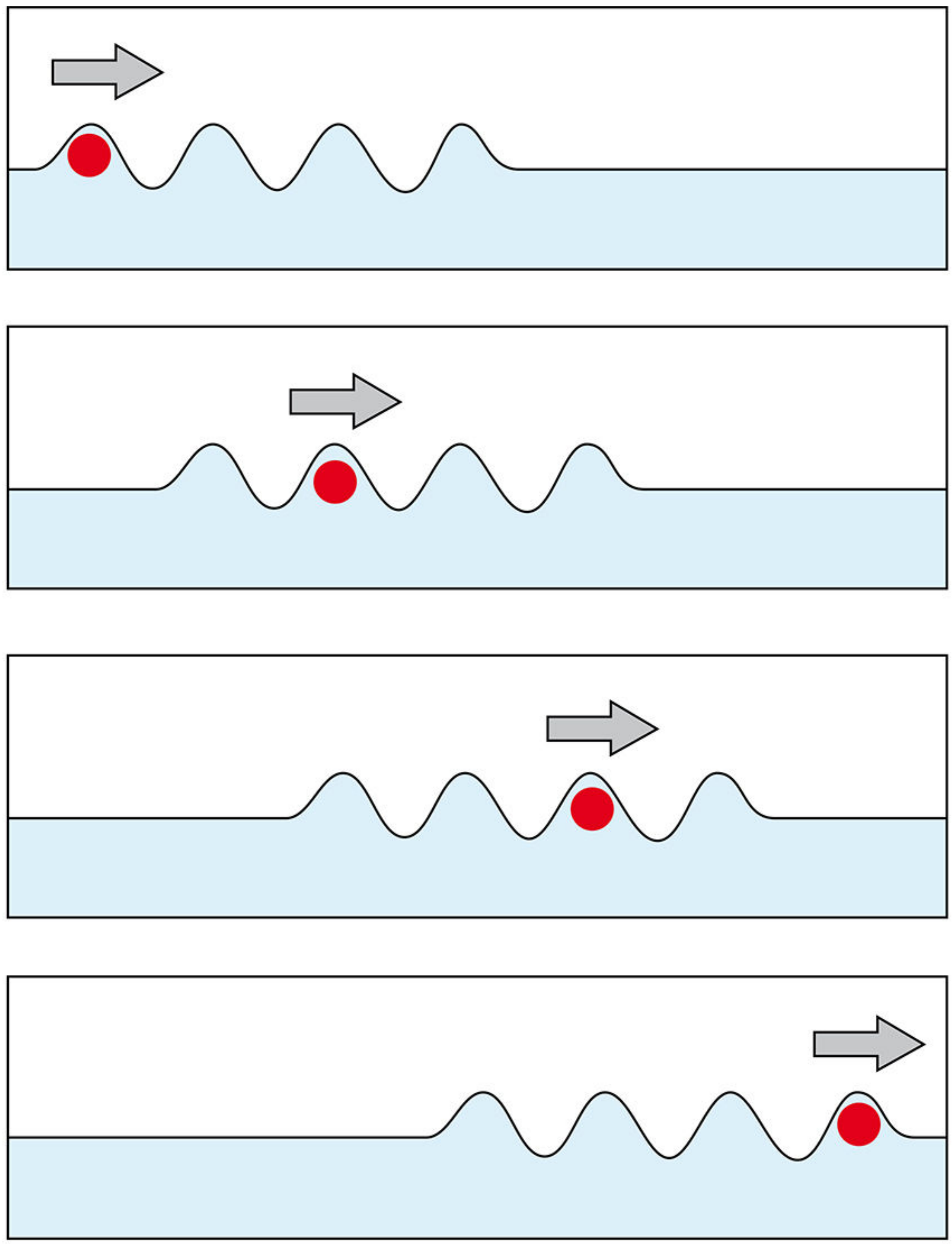

Figure 5.9: Wave speed = twice group speed. Individual waves are born at the back of the group, travel to the front and then disappear.

This principle implies that a single wave cannot propagate very far on its own in the deep ocean. In fact, a single wave does not exist for very long at all. It is the group itself that carries the energy (the message) across the ocean, while individual waves in that group are constantly changing – like a three-piece band that has gone through two drummers, three guitarists and two lead singers – none of the original members remain, but the band still plays the same music.

Earlier in this chapter, I stated that the speed in metres per second of the swell itself in deep water is about 0.78 times the period. This is the group speed, not the individual wave speed. The individual wave speed is twice that value; therefore, the speed in metres per second at which individual waves travel in deep water is approximately 1.56 times the wave period in seconds.

Swell-tracking

Tracking real swells across the ocean, or even just measuring their characteristics as they arrive at some point, gives information that can be used to verify the basic theoretical principles I have been talking about in this chapter.

As already mentioned, swells can propagate immense distances without much energy loss. Deep-ocean swells follow **great-circle routes** around the world. A great circle is the shortest path between two points on a sphere, such as the Earth. In theory, a swell could travel, from its initiation, as far as a great-circle route would allow it before it hit a continent. Probably the longest distance a hypothetical swell could travel would be generated from a low pressure centred, say, just south of South Africa, with strong west-north-west winds on its north-east flank. In theory, this storm could send a swell on a great-circle route all the way to Vancouver Island, Canada, passing Tasmania, tracking through the Tasman Sea, brushing past New Zealand and hitting Fiji and Hawaii on the way. It would be a distance of about 22,000 km – more than half way around the world (Figure 5.10).

Figure 5.10: Great-circle route taken by a hypothetical swell generated off South Africa. The distance travelled would be about 22,000 kilometres.

As far as I know, a swell like that hasn't been measured by anyone, although people have managed to track swells over very large distances. Some of the first swell-tracking experiments around the middle of the twentieth century are now legendary. They proved that oceanic swell propagates along great-circle routes, can travel immense distances without much attenuation and does, in fact, obey the rules of dispersion as predicted theoretically many years earlier.

One of the first documented swell-tracking studies was by Cambridge mathematicians Norman Barber and Fritz Ursell in 1948. They measured swell arriving at Pendeen Lighthouse in Cornwall, England, from a tropical storm on the other side of the Atlantic. The dominant period of the swell started off at about 19 seconds and progressively went down until it reached about 13 seconds three days later. The fact that the long-period waves arrived first and the shorter ones arrived later proved that radial dispersion really does exist, and the longer waves really do travel faster than the shorter ones.

In 1966, another important experiment was undertaken by Walter Munk and a host of scientists from Scripps Institute of Oceanography in California. Six wave-measuring stations were deployed over a 10,000-km stretch on a great-circle route between the Southern Ocean south of New Zealand, and Yakutat, Alaska. Over a period of about two and a half months, twelve major swell events were tracked from station to station. One of the most important findings was that the total wave energy in long-travelled swells is attenuated very little. Once the effects of island blockage and circumferential dispersion had been taken into account, the total energy loss at distances of more than 1,000 km from the storm centre was practically zero.

Since then, the results of those classic experiments have been proven many times. More and more subtle variations have been discovered as instrumentation has improved and global observations have increased. For example, the worldwide network of wave buoys is of great use for detecting and analyzing long-distance swells. And instruments mounted aboard satellites are being used increasingly to track swells, although the data is notoriously difficult to analyze.

6 Waves in Shallow Water

Introduction

In Chapter 5 we talked about what happens to waves as they propagate away from the generating area as free-travelling swell. The observations we made all assumed that the waves were travelling across the deep ocean, where the sea-floor is so far away from the surface that it has no effect on the shape, speed or direction of the waves.

In this chapter we will go a little further, and see what happens to the waves as they come into shallow water where the geometry of the sea-floor – the **bathymetry** – starts to affect them. With a few simple rules about the behaviour of waves, we can begin to understand how a simple swell-line hitting a reef, point or sandbar can change dramatically as it propagates over the changing bathymetry. Thanks to the many different configurations possible, an incoming swell-line can be transformed into many different kinds of waves for surfing.

The main way in which the bathymetry affects a wave approaching the coast is by making it bend, through the process of **refraction**. But also, as the waves come into really shallow water they tend to grow just before they break, especially if they hit a particularly shallow reef.

Wave speed in shallow water

As we saw in Chapter 5, in the deep ocean the orbital motions of the particles beneath the waves diminish to nothing way before they reach the sea-bed (see Figure 5.4). But once the water starts to get shallower, these orbital motions begin to reach all the way down to the bed, which slows down the waves (Figure 6.1).

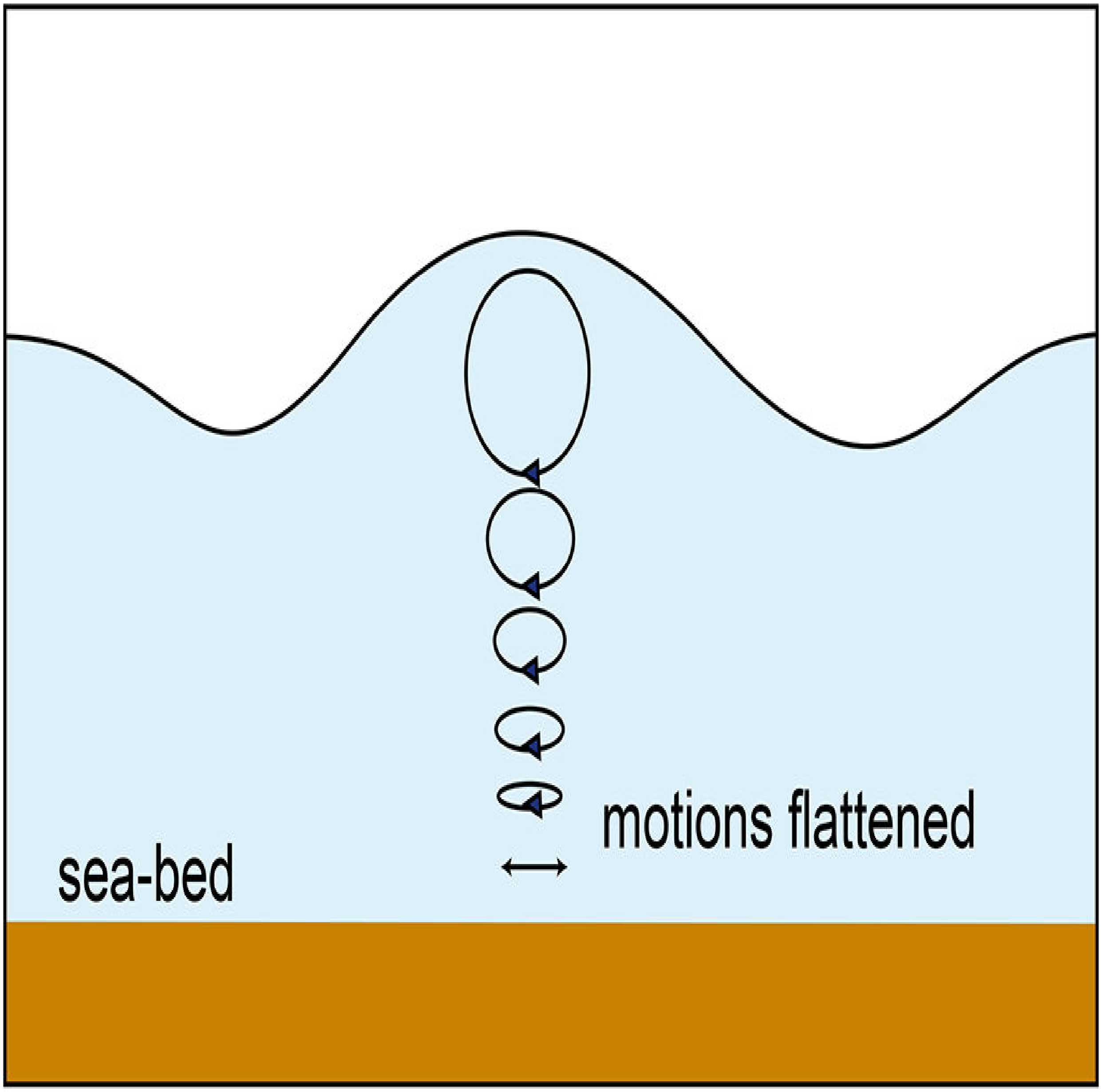

Figure 6.1: In shallow water, the orbital motions below the waves reach right down to the bed.

Just how this happens is a little complicated, but it can be explained in simple terms as follows: The sea-bed is far from flat, especially from the point of view of the water molecules. It has irregularities in it which the molecules bump into as they try to move back and forth along the bed, which slows them down. This change in motion is communicated upwards through the water column in the same way as the original orbital motions are transmitted from the surface downwards – by each molecule 'touching' its neighbour, rather like billiard balls. As a result, the slowing down of the molecules on the bed produces a slowing down of the molecules on the surface, which causes a slowing down of the waves themselves. Curiously, there is very little overall loss of energy.

At what depth does this start to happen? Well, the depth at which the waves start to 'feel' the bottom depends on their wavelength. Longer waves have larger, orbital motions and also, because they travel faster, they carry more energy. Therefore, the orbital motions of longer waves reach down further below the surface, so longer waves 'feel' the bottom before shorter ones. The depth at which the presence of the sea-bed starts to affect the wave speed is about half the deep-water wavelength, or about 0.8 times the period squared, with the period measured in seconds. I prefer to talk in terms of period rather than wavelength, because the period is something we see quoted much more than wavelength and, importantly, the period doesn't change as the waves travel into shallow water (Figure 6.2).

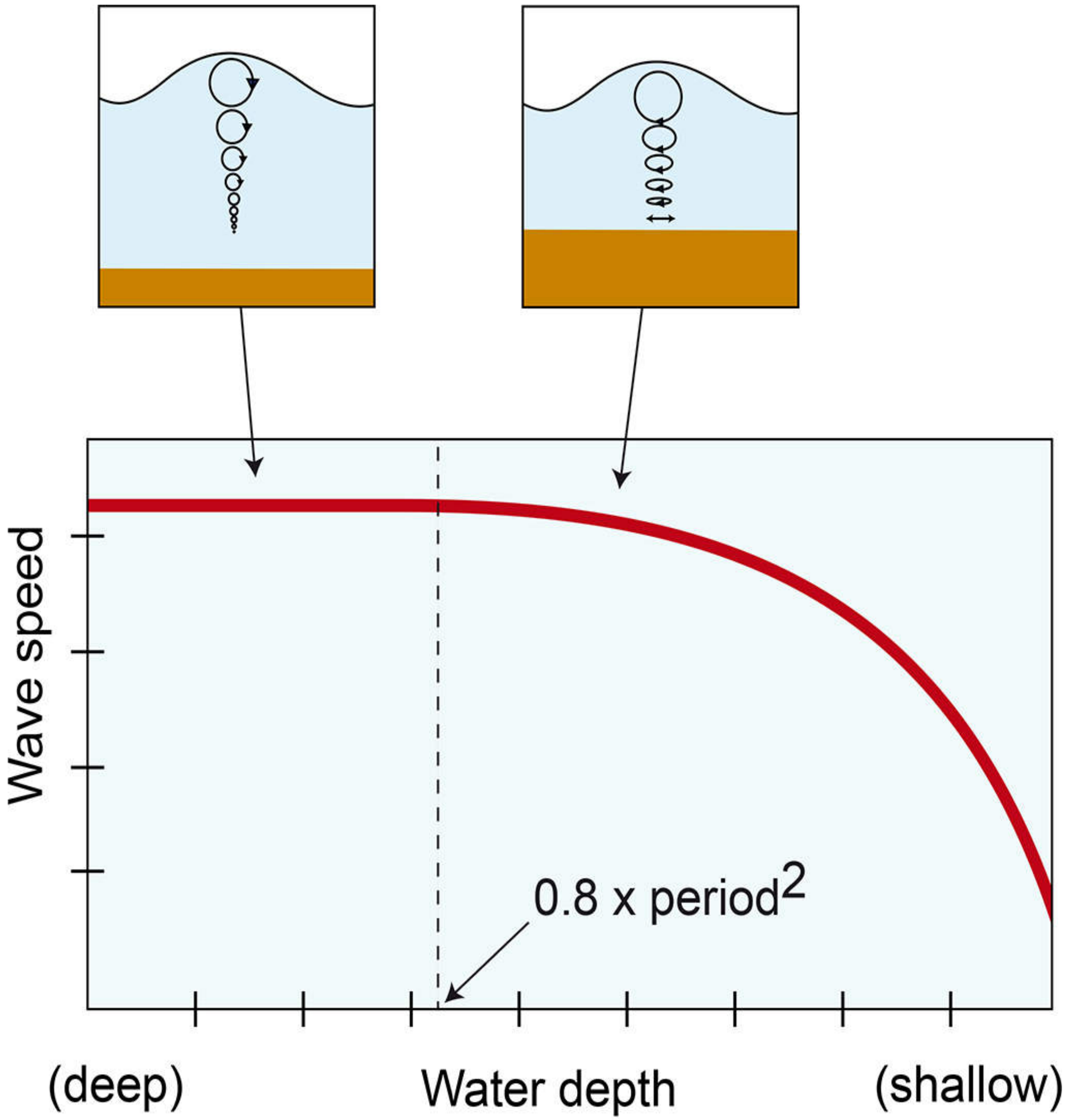

Figure 6.2: Once the depth goes below about 0.8 times the period squared, the wave speed becomes dependent on the depth as well as the period.

In Chapter 5 we stated that, in the deep ocean, the speed of the waves depends only on the wave period. The individual wave speed in metres per second could be calculated by multiplying the period in seconds by 1.56. But once the depth goes below about 0.8 times the period squared, the wave speed must be calculated using the *full dispersion equation*, which takes into account period and water depth. The equation is a little ugly, so I'm not going to go into it here, but it can be looked up in most basic oceanography textbooks.

As the water continues to get shallower, the wave speed gradually loses its dependence on wave period and becomes more dependent on the depth itself. Finally, in really shallow water just before breaking, the wave speed completely loses its dependency on period, and is only dependent on the water depth.

As soon as the depth begins to influence the wave speed, the waves travel at different speeds over different depths. In fact, different parts of the same wave travel at different speeds according to the depth contours. If this happens, the wave bends, or refracts, according to the particular shape of the sea-bed over which it is propagating.

Refraction: basics

Refraction is the bending of a wave front as it travels at different speeds over water of different depths. When one part of a wave travels more slowly than another, the wave bends towards the slower part. Since the part of the wave over shallow water travels more slowly, the wave always bends towards the shallower water. This is easy to visualize if you think of a single swell-line coming in towards a shallow reef next to a deep channel. The part of the swell that finds itself over the reef slows down, whereas the part in the channel keeps going at the original speed. As a result, the wave bends in towards the reef (Figure 6.3).

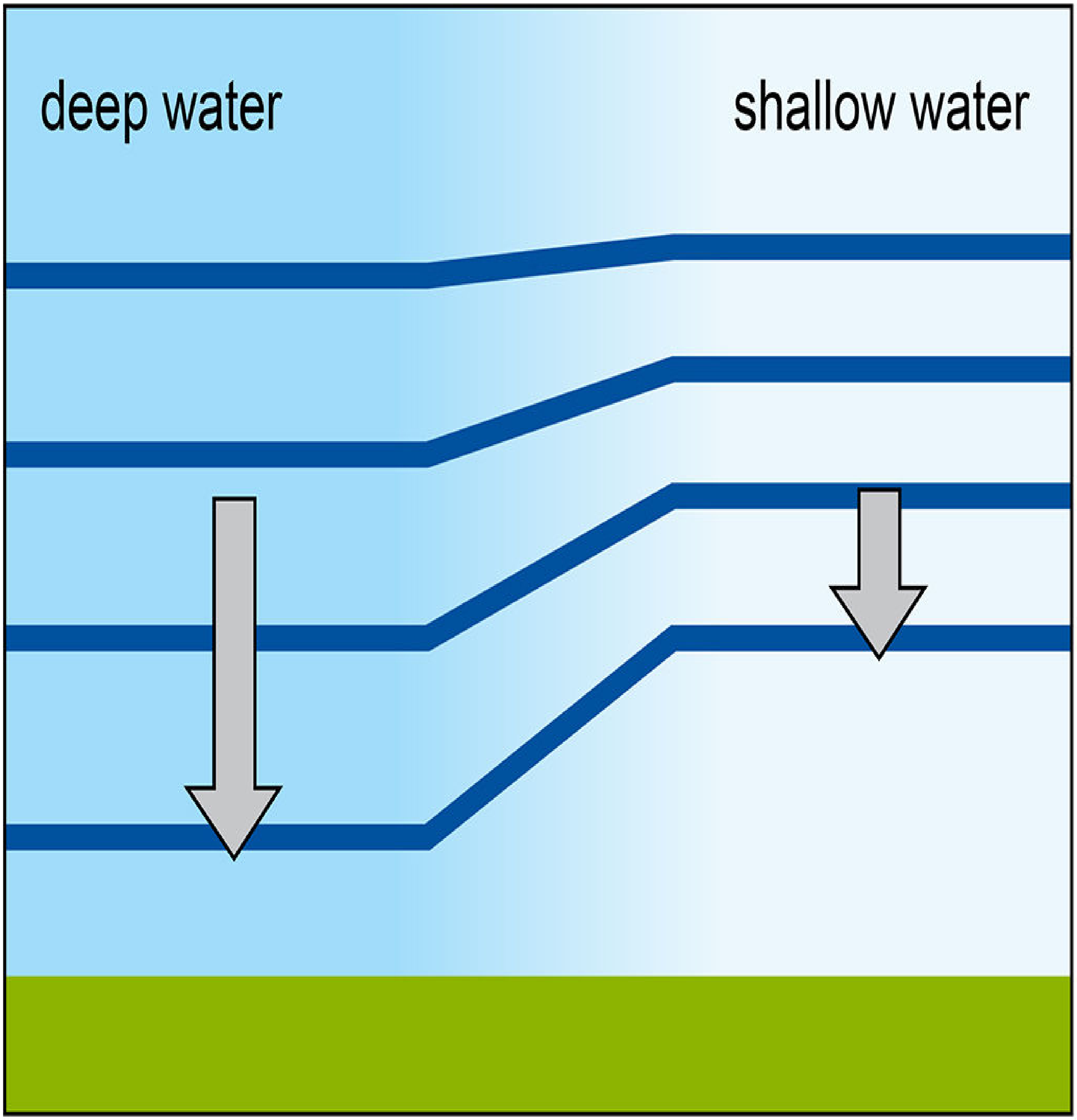

Figure 6.3: The simplest kind of refraction. The wave front slows down over shallow water, therefore it bends towards the shallow water.

A good analogy might be travelling along in a car with the brakes binding on one side. The fact that one side of the car is slowed down relative to the other means that the car veers off to one side; it veers off to the side with the binding breaks.

In reality, the sea-floor bathymetry is never as simple as that shown in Figure 6.3, and waves never all approach from exactly the same angle. So there are many ways in which refraction can change the characteristics of a surf break.

Focusing and defocusing

Now let's consider a slightly more complicated situation. The way that refraction acts upon waves when they hit a shallow area with deep areas either side of it, is different from the way that it acts upon them when they hit a deep area with shallow areas either side of it. In fact, the result can be radically different.

Imagine a straight swell-line approaching a large area of shallow water, but with a relatively deep trench in the middle. The whole swell-line slows down, except for the section in the middle that is propagating over the trench. This part of the wave keeps going at its original speed. As a result, the wave bends outwards away from the deep water, spreading the energy out over a wider area. The middle part of the wave, instead of becoming bigger and more concentrated, becomes weaker and smaller. This kind of refraction is called **convex refraction** or **bathymetric defocusing** (Figure 6.4).

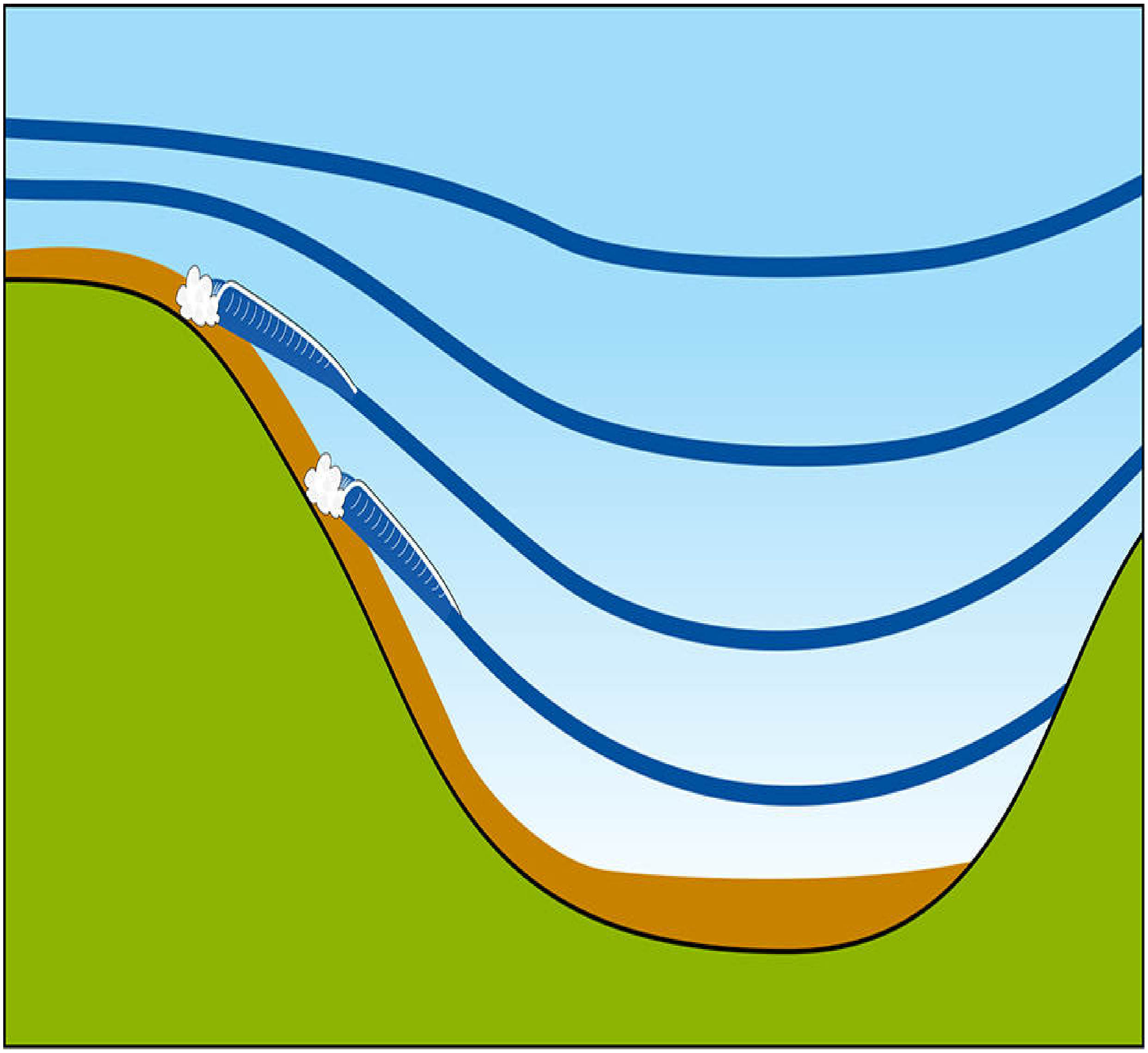

Figure 6.4: Convex refraction. A bay with shallow water either side causes the wave front to bow outwards, its energy being spread over a wider area.

But if the swell-line approaches a shallow slab of reef sticking out from the shore, with relatively deep water either side of it, the opposite happens. The part of the swell-line that propagates over the shallow reef slows down, but the parts either side of it over deep water keep going at their original speed. Therefore, the wave bends inwards from both sides, towards the shallow reef. This concentrates the energy into the middle, turning what was a straight swell-line into a peak. This kind of refraction is called **concave refraction** or **bathymetric focusing** (Figure 6.5).

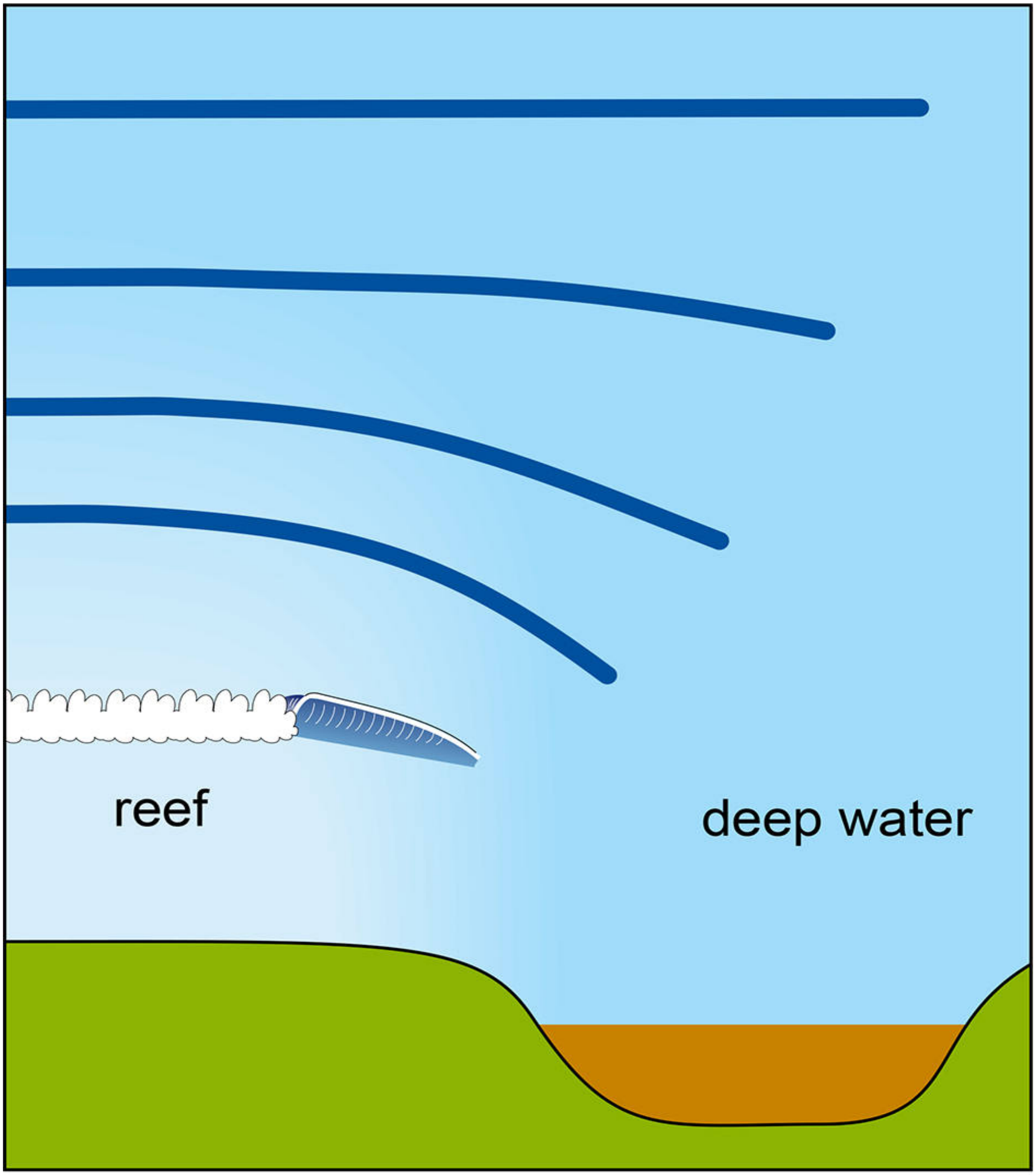

Figure 6.5: Concave refraction. A shallow reef next to a deep channel focuses most of the wave energy on to the reef.

By making the waves bend in a certain way before they break, refraction plays a vital role in determining the nature of a surf spot. For example, in the classic point-break setup you might find a headland with a reef running along the side, next to which is a bay with deeper water. The stretch of coast where the waves break (on the side of the headland) is almost at right angles to the direction of wave approach. As the swell-line approaches, one end of it slows down as it hits the shallow water, causing it to bend in towards the side of the headland while the rest of it continues on towards the beach. The waves are defocused and, because the energy is spread out over a wider area, they are reduced in size. Although a point-break tends to reduce the original size of the waves, it makes them long and walled up, often maintaining their size all the way down the line. One of the most famous examples is Jeffrey's Bay in South Africa, where the wave actually gets bigger for a while after the take off.

Another example of a classic type of break, but with a completely different set of characteristics, is the kind of reef-break where the waves are focused on to a slab of rock sticking out from a stretch of open coast, or off the end of a headland. The headland would typically be next to a deep bay containing a steep beach or even a harbour. The reef is orientated so that one part of the incoming swell-line hits the end of the protruding reef while the rest of it continues on in deep water. The swell-line bends in towards the reef, focusing most of its energy on to the reef itself. The result is a bowling wave where much of the energy that would have gone into the deep area is, instead, concentrated into the peak. This makes the wave bigger and more powerful, but shorter and more concentrated. The reef acts as a 'swell magnet', and the increase in size can sometimes be surprising.

A good example of this kind of break is Sunset Reef in South Africa. Here, the refraction is so pronounced that some days you can find 15-foot waves breaking over the reef while the middle of the bay is virtually flat-calm. The wave itself is short, but much bigger and more powerful than other spots along the same stretch of coastline.

A further kind of break is the beach-break. With a beach-break, refraction on to the existing sandbars dictates where along the beach the waves will break, and how they will break. Most beaches, provided they are of a certain length, contain some kind of undulations in the sand in the alongshore direction. Because of refraction, the waves are focused on to the shallow areas (the bars) and defocused away from the deep areas (the channels). This gives the classic scenario of a beach-break with various peaks along its length and channels in between (Figure 6.6).

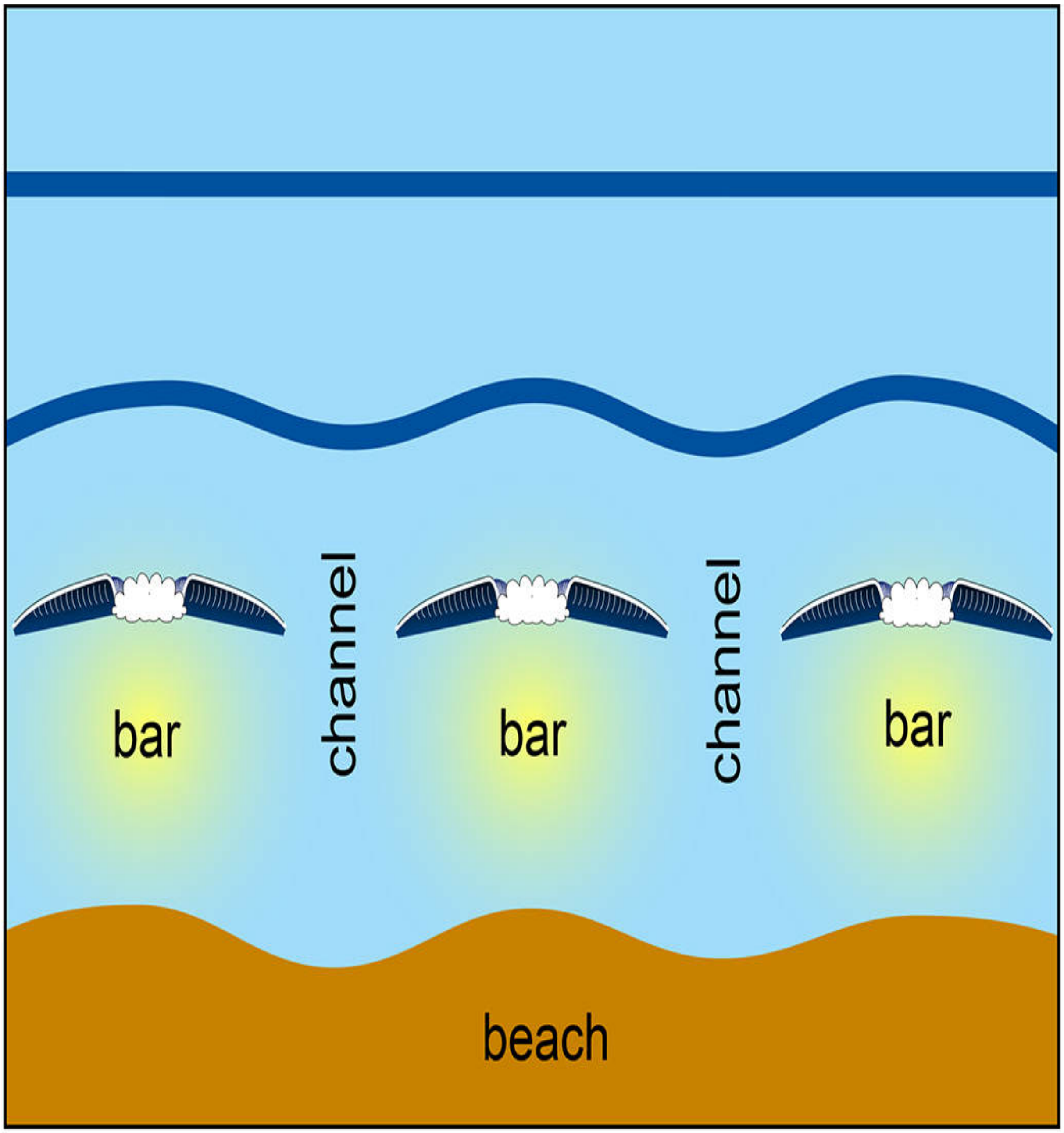

Figure 6.6: Beach-break with sandbars along its length. Concave and convex refraction produce a series of peaks with channels in between.

Of course, not all beach-breaks are like this. For example, on a small ‘pocket’ beach there might be defocusing around both headlands, which slows down the two extremes of the wave front while the middle keeps going at its original speed, rather like a double point-break but without the wave breaking along the side of the headland. This kind of refraction causes the wave front to approach the shore almost parallel to the depth contours themselves. The wave front is ‘bowed out’ as it approaches the typically half-moon shaped beach. Unless there is some kind of sandbar or reef in the middle of the beach, the wave will close out (Figure 6.7).

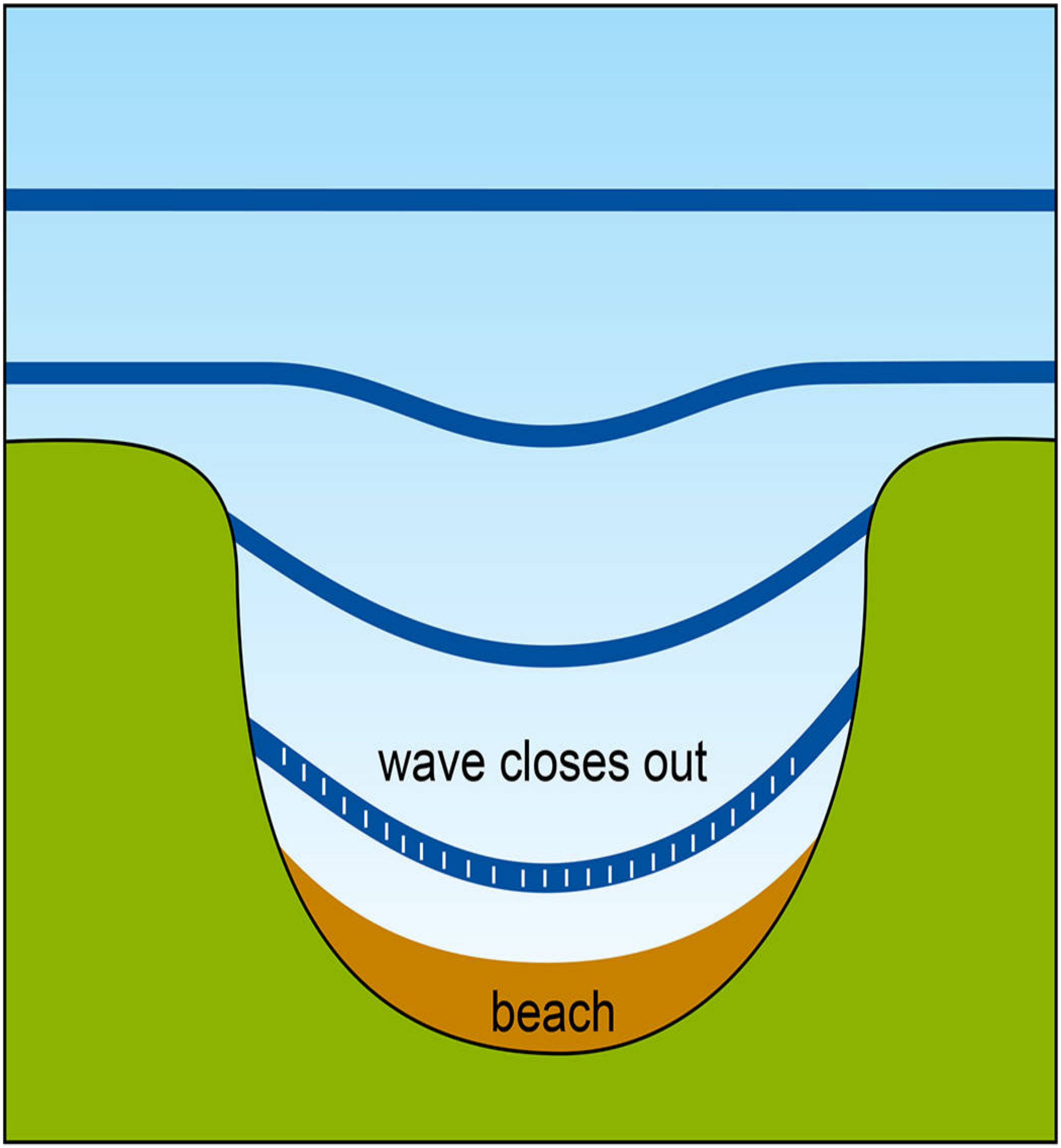

Figure 6.7: On a small 'pocket' beach, the wave front follows the depth contours and eventually closes out.

Longer waves bend more

An interesting property of refraction is that it is period-dependent: the longer the period, the more the refraction. Or, put another way, the longer waves begin to refract in deeper water than the shorter ones because they 'feel' the bottom first. Referring back to Figure 6.2, we can see that increasing the period makes the wave speed start to depend on the depth earlier (increasing the period makes the vertical dotted line move towards the left of the diagram). When wave speed starts to depend on depth, refraction can occur.

One implication of this is that a high-quality swell, which typically contains a higher proportion of longer period waves, will respond to the effects of refraction more readily than a low-quality swell. This is also true for the first waves of a new swell, which are the longest ones. It is often the case that, at the beginning of a swell, a spot with bathymetric focusing has much bigger and more powerful waves than other spots along the same stretch of coast. In contrast, towards the end of the swell, this difference would be much less noticeable.

Also, the quality of the waves at some point-breaks is affected by the period; the longer waves at the beginning of a swell might 'hook in' better because they are refracted earlier, whereas the shorter ones towards the end of the swell might end up having nasty sections or going 'fat' into the middle of the bay. It all depends on the particular characteristics of the spot.

Wave height transformation

As waves start to move into really shallow water, just before they break, they tend to increase in size. This is a separate mechanism from refraction, but just as important. If the transition from deep to shallow water is quite sudden – if the waves hit a shallow reef out of deep water, for example – the wave height increase can be quite noticeable. If you then combine it with the effects of concave refraction as described above, the wave height increase can be phenomenal.

How does it work? Imagine a wave train containing several waves moving from already quite shallow water into even shallower water. In really shallow water, the wave speed depends on the water depth, so the first waves in the train slow down first. A side-view snapshot of the wave train would show the waves at the front of the train travelling more slowly than the ones at the back. As a result, the waves are squashed up together like an accordion (Figure 6.8).

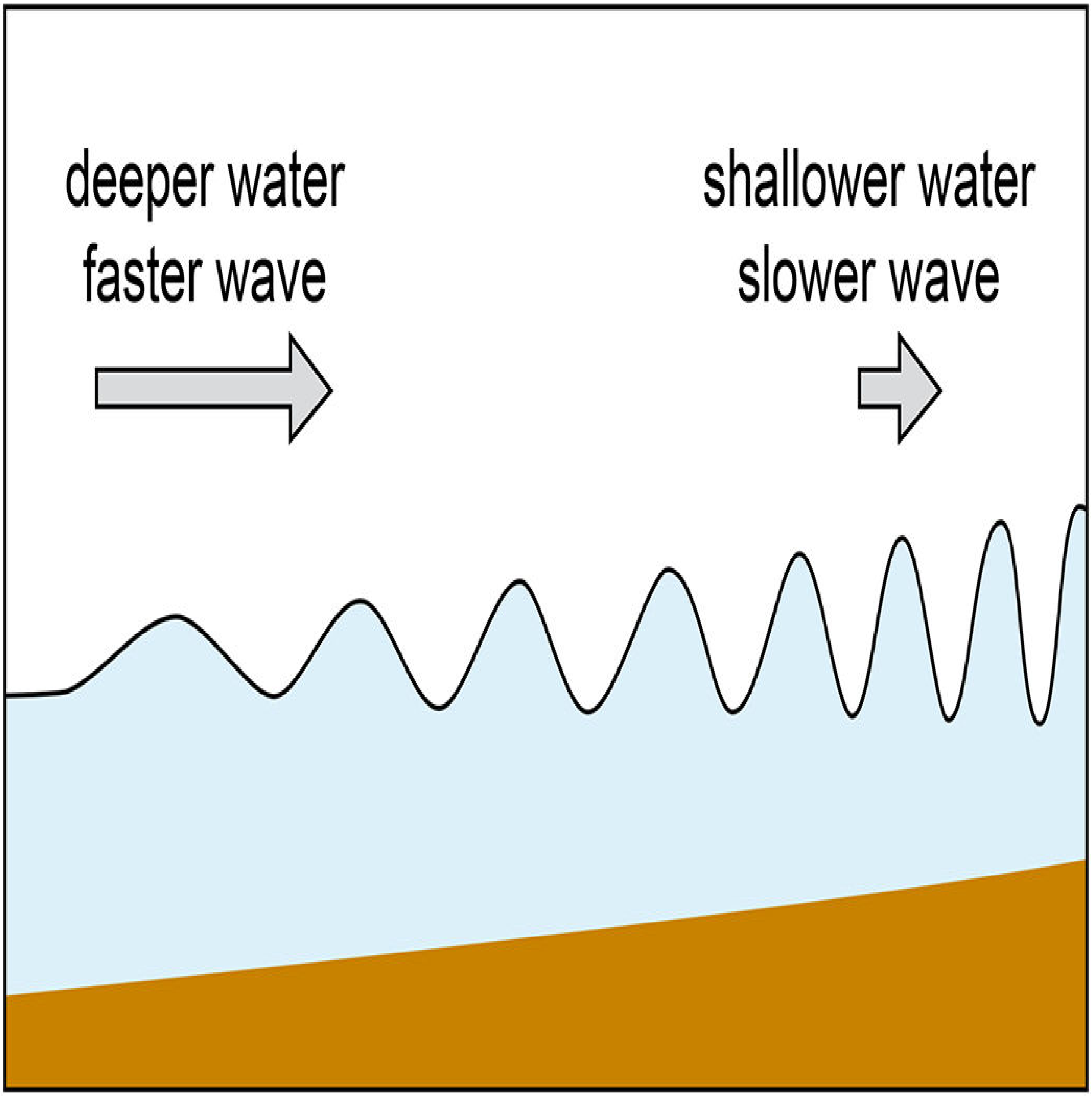

Figure 6.8: As a wave train approaches shallow water, the waves at the front slow down, making them squash up and become higher.

As they squash up, they get higher. This is because the same amount of energy is concentrated into a shorter distance. If you can't imagine this, take a coat-hanger and bend it so that it looks like a wave-train containing several waves. Then push the wave-train from the back while keeping the front stationary. The waves in the coat hanger will get higher as they are pushed closer together.

One important thing to remember is that, as the waves come into shallow water, the wave period always remains the same. The distance between one wave and the next (the wavelength) reduces by exactly the right amount to compensate for the reduction in speed as the waves hit shallow water. As a result, the time taken for each wave to pass a given point (the period) never changes.

7 Wave-breaking

Introduction

In Chapter 6 we looked at how the shape of the sea-floor causes a wave to bend and deform as it comes into shallow water, which can radically affect the way it eventually breaks. In this chapter we will look into the wave-breaking process itself – why waves break and when they break, and the important differences between the ways in which the ocean finally unloads its energy on to a reef, point or sandbar.

Different kinds of breaking waves are suited to different styles of surfing. For example, mellow, slow breakers might suit the beginner or the laid-back longboarder who doesn't want anything too radical. These waves are easy to take off on, easy to ride and easy to manoeuvre on, but might be considered boring by the more radical surfer. The faster, hollower wave, with the possibility of a tube, is more exciting, but, at the same time, more risky and difficult to handle. The fastest and heaviest waves are the most exciting, but eventually reach the stage where they become impractical to anyone but the best surfers in the world.

Each of these kinds of waves has a different **breaking profile** – the shape of the wave looking side-on. The profile is dependent upon various factors, such as the height and wavelength of the wave itself, and the strength and direction of the wind. But probably the most influential factor is the shape of the sea-floor. For example, most waves that break on shallow slabs of granite are almost invariably steeper and more difficult to surf than waves that break on gently-sloping beaches, no matter what the wavelength of the wave is, or what direction the wind is blowing.

We will start by looking into why waves break in the first place, and in what water depth they are theoretically supposed to break. Then we will describe how the breaking depth can vary according to the sea-floor, the kind of wave and the wind, and how different breaking depths can mean very different wave profiles.

Why do waves break?

In Chapter 6 we saw that, when waves start to propagate over shallow water, they slow down and bend through the process of refraction. When they come into even shallower water, further changes take place, and eventually they break. But why do they break? The reasons are many and complex, but here is one very simple explanation.

If the depth of the water is less than about one-twentieth of the wavelength of the wave, the speed of the wave depends only on the depth of the water. The exact relation between wave speed and water depth (wave speed in metres per second = 3.13 times the square root of the depth in metres) was published by G. B. Airy in about 1845, along with a set of equations relating to water motion in and around a wave.

Airy wave theory can be used successfully when the height of the wave is considered negligible compared with the water depth. However, in *really* shallow water, this assumption is no longer valid. When the wave comes into really shallow water, just before it breaks, wave height increases while water depth decreases, so the height of the wave is no longer insignificant compared with the water depth. In other words, the bottom of the wave is significantly nearer the sea-bed than the top of the wave. Effectively, the bottom part of the wave is in shallower water than the top part of the wave. This means that the slowing-down influence of the bed will be greater on the bottom of the wave than on the top.

As a result, the bottom of the wave slows down more than the top of the wave. This effect becomes progressively more pronounced as the water gets shallower, eventually reaching a point when the top of the wave overtakes the bottom, making it spill forward and break. The wave is ‘tripped up’, just the same as if someone trips you up as you are walking along – the bottom half of you begins to travel more slowly than the top half, so you fall over head first (Figure 7.1).

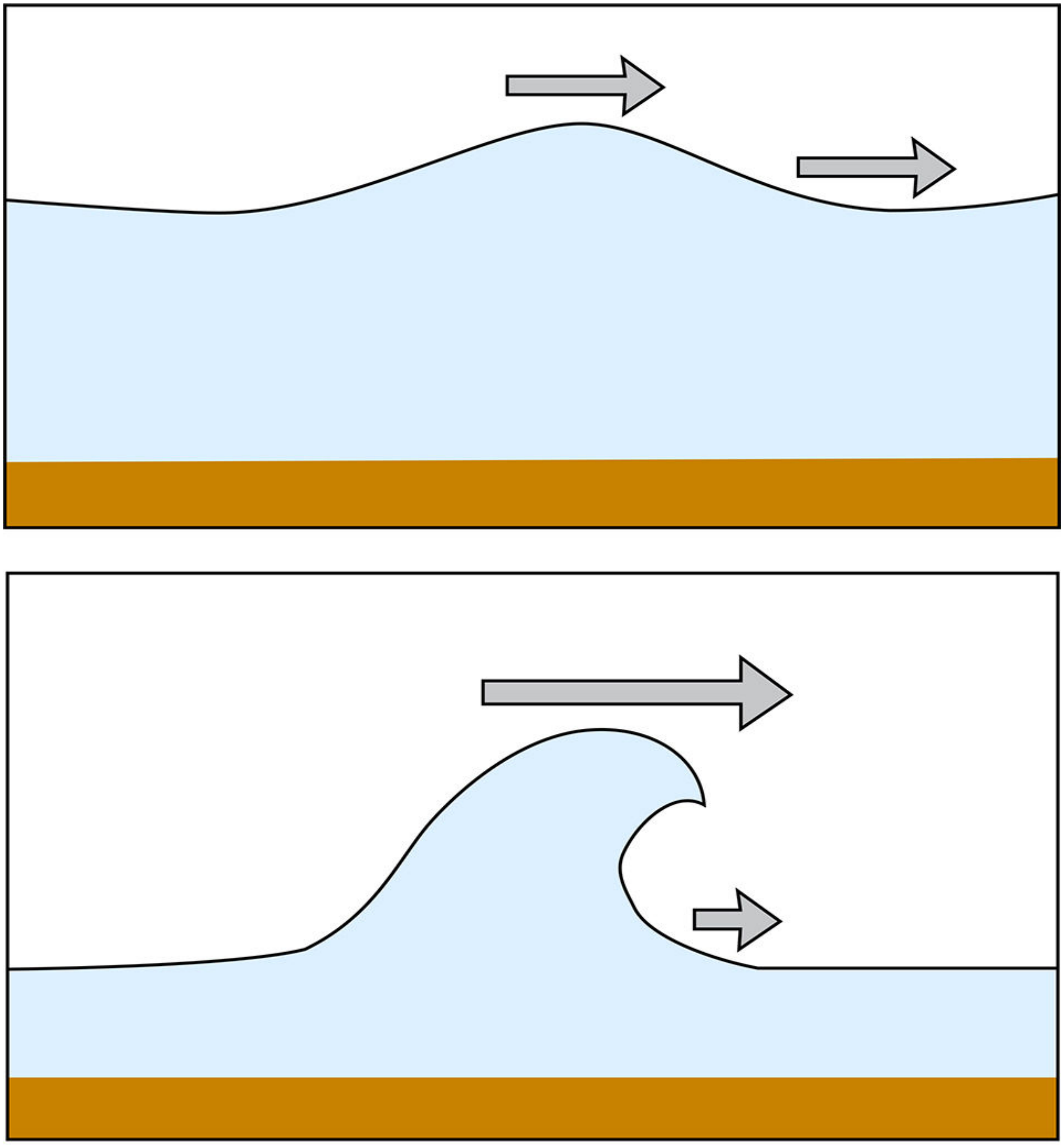

Figure 7.1: In deep water, the top and bottom of the wave travel at about the same velocity, but in shallow water the top travels faster than the bottom.

When will a wave break?

How can you tell when a wave will actually break? In other words, at what depth of water does a typical wave break? Through years of experiments, mostly in laboratories, experts have come up with an average depth of about 1.3 times the height of the wave. That means that a typical 1-m wave will break in about 1.3 m of water.

However, it's not quite that simple. As surfers, we all know that the depth under a breaking wave can vary enormously. At a hollow reefbreak on an offshore day, you might find 2-m waves breaking in less than 1 m of water. Or on a very gently sloping beach on an onshore day, you might find 1-m waves breaking in 2 m of water. So the rule is almost totally academic, and can only be used as a guide to the average depth of water in which a wave breaks under ideal conditions.

Importantly, it is the departure of the breaking depth from this average value that governs the wave profile. If circumstances dictate that the wave breaks in water whose depth is less than 1.3 times its height, then its profile will be steeper and more hollow. But if the wave breaks in water whose depth is more than 1.3 times its height, it will be fatter and more gently sloping.

What factors can make a wave break in different depths? One thing that the breaking depth – and hence the wave profile – depends upon, is the suddenness with which the depth changes as the wave propagates towards the shore. If the transition from deep to shallow water is quite gradual, the wave tends to break when the ratio between water depth and wave is near to its theoretical value of 1.3. It breaks at the moment when its top is travelling just that little bit faster than its bottom, just enough for the wave to spill over. The resulting wave profile is quite a slow, gently breaking wave.

In contrast, if the transition from deep to shallow water is very sudden, for example over an offshore reef or 'slab', then the wave finds itself momentarily over very shallow water without having broken. The **shoaling** effect of the reef makes the wave suddenly increase in size at the same time as the water depth suddenly decreases, so the height of the wave is relatively large for the depth of water. The water depth is already less than 1.3 times the wave height, even before the wave breaks. So the speed of the top of the wave is much greater than the speed of the bottom, which causes it to throw out as it breaks, making it steep, hollow and fast (Figure 7.2).

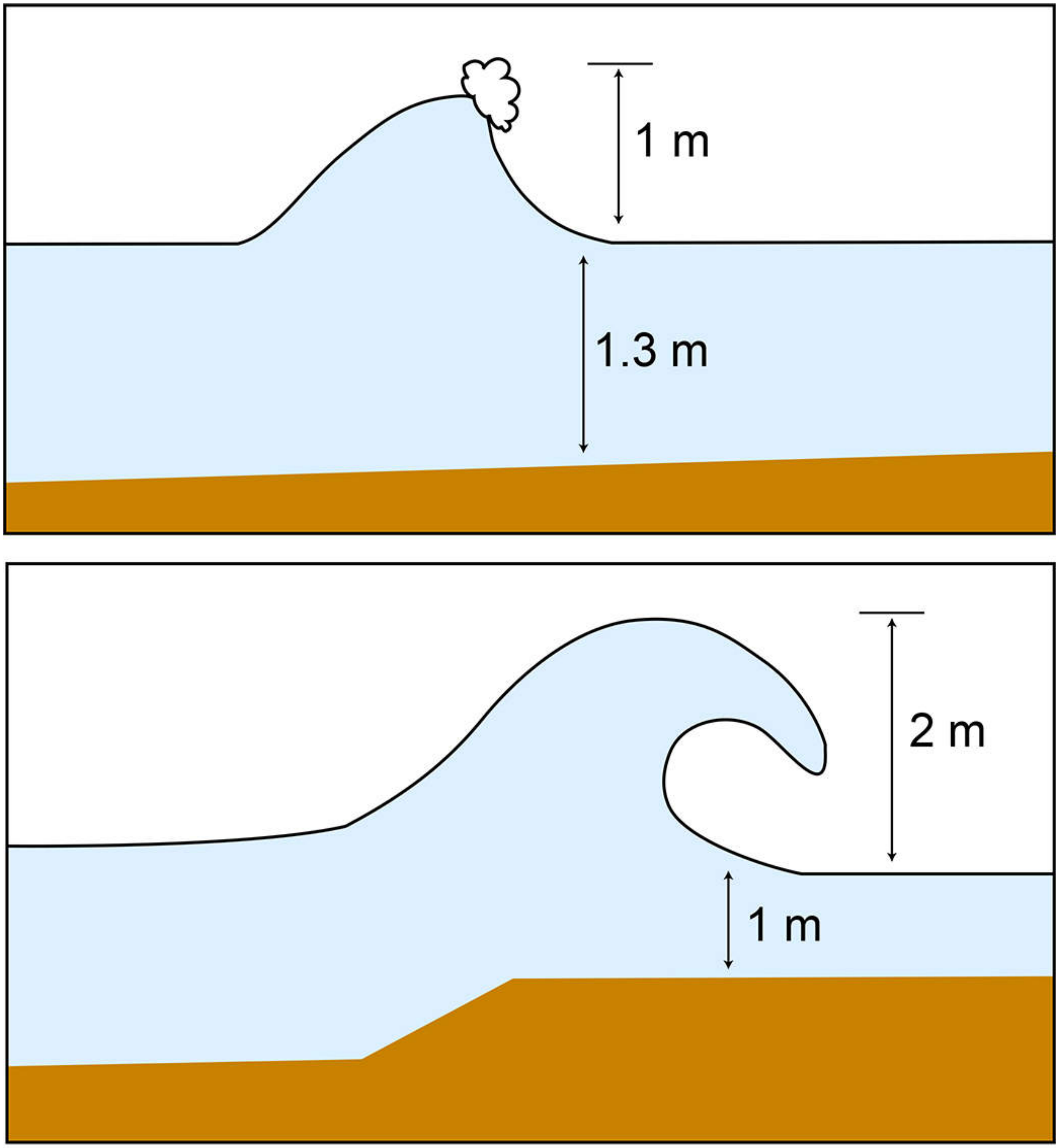

Figure 7.2: In theory, a wave will break when the depth is about 1.3 times the wave height, but if the depth transition is abrupt it will break in shallower water.

Other factors that influence the wave profile

The wave profile depends mainly upon the abruptness of the deep-to-shallow water transition. But it is also influenced by the speed at which the wave comes out of deep water, the **steepness** of the wave before it breaks, whether the wind is onshore or offshore and how strong that wind is. These factors are somewhat interdependent, which complicates things a little more. They are explained below, and illustrated in Figure 7.3.

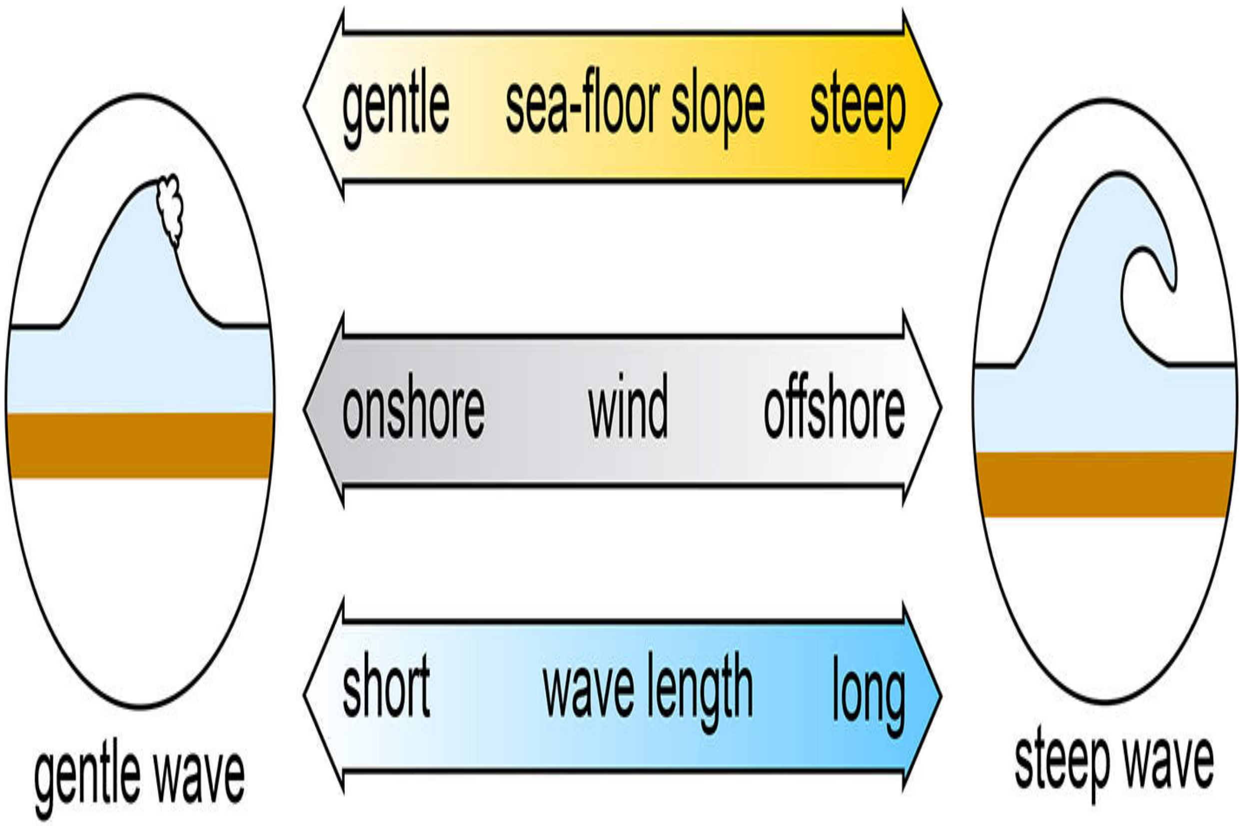

Figure 7.3: Illustration of how the wave shape depends on the slope of the sea-floor, the direction and strength of the wind and the wavelength of the wave.

First is swell quality. The incoming swell might contain a greater proportion of shorter-period or longer-period waves, depending on how far it has travelled, or whether it is a new or old swell. Shorter-period waves also have shorter wavelengths, and shorter-wavelength waves are steeper than longer ones of the same height (longer-wavelength waves are more 'stretched out' and shorter-wavelength waves more 'squashed together'). Therefore, shorter-wavelength waves tend to break in deeper water.

Waves from a local storm or an old dying swell tend to break further out, in deeper water. This makes them relatively slow and mellow. In contrast, waves from a distant storm, or the first waves in a growing swell, tend to be further apart and less steep because they have a longer wavelength. They therefore need to travel a greater distance over shallow water before they become steep enough to break. When they do eventually break, they unload a greater amount of energy in a shorter distance, so they are thicker and more powerful.

The wind is another factor. An offshore wind physically holds up the wave, stopping it breaking until it gets into shallower water. That is why there are more tubes when the wind is offshore. An offshore wind also 'cleans up' the swell itself, removing the short-period windsea and leaving the long-period swell waves. Because long-period waves break in shallower water anyway, an offshore wind has a double effect in making the waves steeper and more tubular.

A strong onshore wind also has a double effect. It causes the waves to spill over early, making them break in deeper water, but also locally generating more short-period windsea, which tends to break further out anyway. With an onshore wind, the waves are guaranteed to break more slowly and with less power.

'Official' breaking wave profiles

Knowing the profile of the wave you are about to surf is pretty important. Through a series of experiments involving miniature waves in special laboratories, experts have quantified 'breaker types' into three, or sometimes four, different categories. They are:

… the **spilling breaker** – a wave that breaks very slowly with little power and no tube;
… the **plunging breaker** – more powerful and generally the best for surfing, with a possible tube;
… the **surging breaker** – this just surges up and down the beach without breaking in the normal sense, so isn't really any use for surfing.

In some textbooks, there is a fourth category of wave between plunging and surging, called the collapsing breaker.

These breaker types are categorized in terms of two of the factors described above, which determine in what water depth the wave will break – namely beach slope and wave steepness. Non-steep (swell) waves on a steep beach tend towards the ‘surging’ type; steep (wind) waves on a gently sloping beach tend towards the ‘spilling’ type. But this standardized theory takes no account of the effect of wind, which, as pointed out above, is actually very important.

8 The Temperature of the Water

Introduction

The temperature of the water is a very important factor for surfing. In some parts of the world you must have adequate protection from the cold water, or you will miss the best surf of the year. The temperature may even keep some people out of the water altogether – sometimes even the best waves are not worth the pain of ice-cream headaches, numb hands and feet, and the hassle of having to enter the water Michelin-Man like, with boots, gloves, hood and several millimetres of neoprene.

In this chapter we will look at what controls the temperature of our surfing waters. In some places, water temperature changes radically from season to season; in others it hardly varies throughout the year. Changes in water temperature are not completely controlled by latitude or season. So we will examine some of the factors that completely override any temperature differences due to latitude and season, which go some way towards explaining why water temperature seems to have a mind of its own.

Radical and not-so-radical temperature variations

Around the world there are places where, at first glance, you would think the water would be so cold that it would be madness even to contemplate surfing. But in some of these places the water is actually quite warm. There are other places where the air temperature gets blazing hot during summer, but the water remains freezing cold. Sometimes, water temperature doesn't seem to relate in any way to latitude, or to whether it is summer or winter. In some places, water temperature remains remarkably constant throughout the year; in others, it goes from one extreme to the other.

If you go towards the poles, logically you would expect the water to get colder in winter because the solar radiation hitting the surface of the Earth is much less than it is in summer. But sometimes this does not happen. In Europe, some of the most popular surfing areas have quite unexpected water temperatures. For example, in France in early September you can happily surf without a wetsuit, whereas in Portugal, which is nearer the Equator, the water is much colder. In Portugal, you sometimes need a steamer and boots in early September.

There are physical mechanisms apart from the simple local heating of the water by the Sun's radiation, which have a major say in controlling the temperature of the water. Among these are (a) the relatively slow heating up and cooling down of seawater compared with land, (b) **coastal upwelling**, which regulates coastal waters by constantly replacing them with cold water from underneath and (c) ocean currents, which also regulate coastal waters by constantly replacing them with water from some other part of the ocean. The entire ocean-atmosphere system of the planet is all connected together, so these factors are interlinked. I will explain a little about them below, one by one.

Land-sea influence and specific heat capacity

If you compare mid-ocean islands with inland water bodies, you will find a great difference in temperature variability throughout the year. On most ocean islands, water temperature changes very little from summer to winter, even on islands such as the Azores that are located at latitudes where land temperatures can vary greatly. But on inland water bodies such as the Great Lakes (in north-eastern North America), water temperature changes dramatically from one season to the next. Here it can be over 20°C in summer but pitch to zero degrees in winter.

Ocean islands are small pieces of land surrounded by a vast sea, and inland water bodies are small areas of water surrounded by vast areas of land. One is an inside-out version of the other, if you like. The temperature of the coastal waters of small islands like the Azores is controlled by the surrounding ocean, without any influence from the relatively tiny amount of land that makes up the island. In contrast, the temperature of the coastal waters of inland water bodies such as the Great Lakes is controlled by the huge land mass surrounding the relatively small amount of water. To understand why they behave so differently, we need to go back to the concept of **specific heat capacity** (**SHC**), as mentioned in Chapter 2.

> *The specific heat capacity (SHC) of a substance is the amount of energy in joules it takes to raise the temperature of one kilogram of that substance by one degree Celsius.*

The SHC of a substance is the ability of that substance to store heat energy inside it. Water is much better at storing heat energy than land – in fact the SHC of seawater is about five times as much as the SHC of typical substances that make up coastal land. So, to raise the temperature of an equal mass of land and sea by equal amounts requires about five times as much solar radiation to be pumped into the sea as into the land.

As a result, the seasonal variations in solar radiation have a much greater effect on land temperatures than on sea temperatures. Seasonal temperature swings of the oceans as a whole are much smaller than those of the continents. Therefore, on small oceanic islands, the seasonal temperature variations of the coastal waters follow the relatively small temperature swings of the surrounding ocean; but on inland water bodies, the seasonal temperature variations follow the much larger swings of the surrounding land.

Hysteresis

The particularly high SHC of seawater also means that it takes a long time to heat up and cool down. To raise the temperature of the sea by one degree requires a lot of solar radiation being pumped into it, which takes a long time. Likewise, for the temperature to fall by one degree, the sea has to lose all that energy, which also takes a long time. The delayed reaction to the input and output of energy to and from a system is known as **hysteresis**. It can be seen everywhere in science and engineering, from the changing shape of sediments on the shoreline to the power-steering mechanism of your car. The delayed heating up and cooling down of the sea due to the input and output of solar radiation is a classic case of hysteresis.

This delay also means that seasonal variations in sea temperature tend to be out of phase with the seasons themselves. Maximum and minimum sea temperatures do not occur mid-summer and mid-winter – they occur up to one or two months later. In early spring, for example, just when the weather is starting to warm up, the sea might still be cooling down. Then, in early autumn, just when the air is starting to become a little chilly, the sea might still be warming up (Figure 8.1).

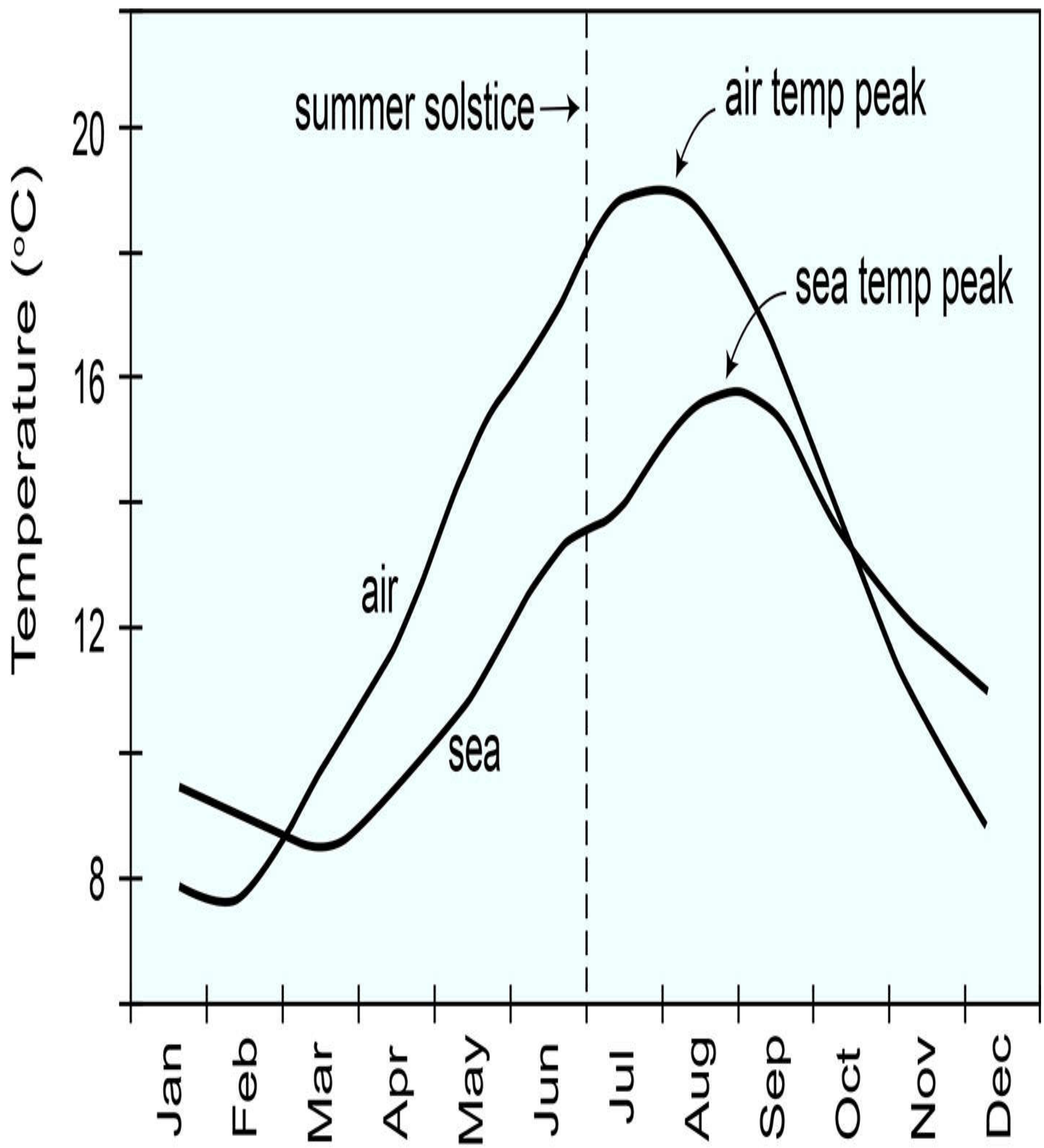

Figure 8.1: Monthly averages of maximum daily air temperatures and monthly average sea temperatures for Newquay, England.

Upwelling

Another reason why coastal water temperatures don’t just behave according to the season or the latitude is a phenomenon called **coastal upwelling**. This is the continual uprising of cold water from underneath to replace the warm surface water, while the surface water is continually removed by strong **trade-winds** – consistent winds that blow on the Equator-ward flank of high pressures, i.e. from the north-east in the Northern hemisphere and from the south-east in the Southern hemisphere. Because the surface water is constantly being replenished, upwelling not only makes the coastal water colder than normal, it also makes it more resistant to seasonal changes.

To understand how upwelling works, we will assume, initially, that the ocean is **thermally stratified**. That means the warmer, lighter water is sitting on top of the colder, heavier water. If there is a strong enough offshore wind, this will blow the surface water away from the coast, allowing the colder water underneath to rise up and take its place. As long as the offshore wind keeps blowing, the water will continue in this cycle: cold water reaches the surface, is blown offshore, and is replaced by more water from underneath. The Sun never gets a chance to heat up the coastal water, so it stays cold (Figure 8.2, lower panel).

You would imagine that a straight offshore wind would make the water move straight offshore. But this is not quite the case. Thanks to the Coriolis force (see Chapter 2), the surface water being driven by the wind does not move in quite the same direction as the wind. It is deflected approximately 45° to the right in the Northern hemisphere, and to the left in the Southern hemisphere. Therefore, a wind blowing from a direction somewhere between side-shore and side-offshore would result in the surface water being driven straight offshore (Figure 8.2, upper panel).

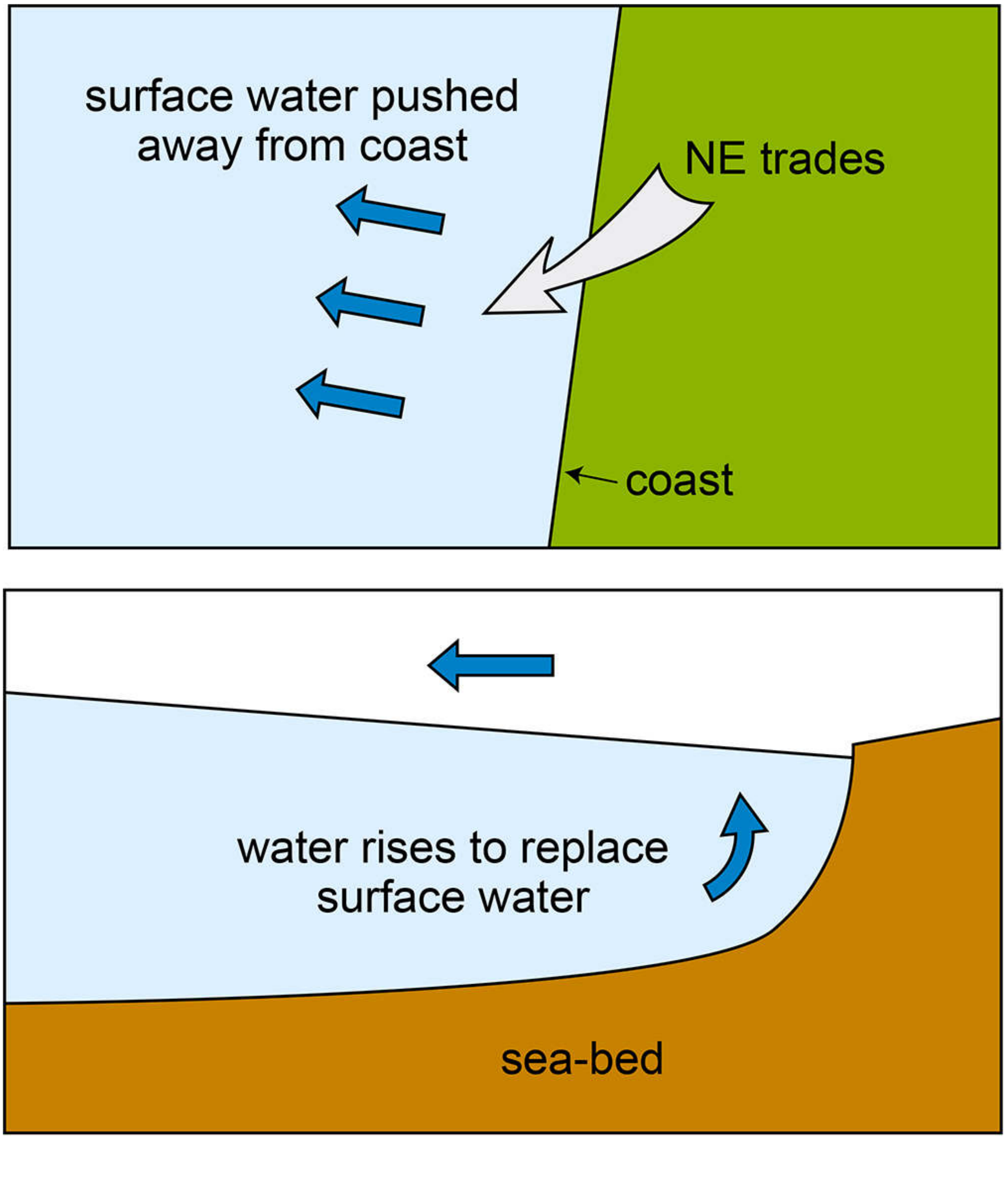

Figure 8.2: Plan view (top) and side view (bottom) of upwelling in the Northern hemisphere.

The amount of upwelling depends principally on the strength of the wind, but this is also a little more complicated than you might first think. The water motion is proportional to the windspeed squared, which means that for every doubling of the wind speed, the water is pushed four times as hard. As a result, sudden relatively small fluctuations in the wind cause disproportionately large amounts of upwelling.

Where and when does upwelling happen? It can occur anywhere the wind is strong enough and in the right direction to cause offshore movement of surface water. The principal upwelling zones of the world are on the western sides of continents where there are strong winds on the eastern flank of a large oceanic high-pressure system. These are north or north-easterlies in the Northern hemisphere, and south or south-easterlies in the Southern hemisphere. In summer, when the high pressure is more intense, the winds are stronger and the upwelling is more noticeable. Typical upwelling areas are Peru, Chile, Portugal, south-west and north-west Africa, and the central west coast of North America. Notice that these are all really good surfing areas too.

A classic example of upwelling is around the south-west tip of Africa, near Cape Town. In winter, the water temperature is normally around 15°C. However, mostly during the summer months, there is a strong south-east wind called the **Cape Doctor**, which blows semi-side-shore on the west coast. As soon as the Cape Doctor starts to blow, the upwelling process begins. By the third or fourth day, the water temperature might have dropped to as low as 9°C. In summer, it is not uncommon to find land temperatures soaring above 40°C at the same time.

Consequences of water temperature for surfing

Finally, we will look at how the temperature of the water affects our experience while surfing, focusing on the effect it has on our perception of the waves themselves.

We have all probably agreed at some time or other that summer swells just don’t seem to have the same punch as winter ones – winter swells are thicker and gnarlier. While wiping out in warm water is almost a pleasant experience, wiping out in cold water is anything but, as wave after icy-green wave slams down on you like a ton of bricks.

There are several possible reasons for this. Perhaps summer swells come from weaker storms, and are therefore somehow weaker; or perhaps the temperature of the water in the generating area is warmer, which in some way makes the waves have less power. But maybe the water temperature itself simply makes winter swells 'heavier' than summer ones because the water is that much denser. Apart from between zero and 4°C, water becomes denser as you decrease its temperature, so a cold-water wave of the same size must weigh slightly more than a warm-water one, which means it must pack a bigger punch, right?

To find out the answer to that question, we need to know how much more energy a cold wave contains than a warm one. The kinetic energy of a wave is proportional to its density times its volume times the square of its velocity. If we compare two hypothetical waves of the same volume and velocity, then the difference in energy just becomes proportional to the difference in density. So all we need to know is the densities of seawater at two different temperatures.

In Atlantic City on the east coast of North America, the water varies from 2°C in winter to 22°C in summer. The density of seawater is typically about 1,028 kg per cubic metre at 2°C, and 1,024 kg per cubic metre at 22°C – a difference of about 0.4 per cent. So a wave in Atlantic City weighs about 0.4 per cent more in winter than in summer. If the wave is travelling at the same speed, it will contain about 0.4 per cent more energy. That is not very much, and it is difficult to imagine noticing the difference when a wave breaks on your head. The volume of water falling on your head when you get caught by the lip of, say, a one-metre wave, is probably about two cubic metres, which weighs about two tonnes. With a couple of tonnes of water coming down on you, an extra 4 kg is not going to make much difference.

But what really makes cold waves 'heavier' than warm ones is the effect the cold water has on the human body. Cold water not only feels uncomfortable, but can quickly cause the body to stop functioning. The fact that water conducts heat away 32 times more rapidly than air, means that submersion in cold water can cool the core temperature very quickly. In cold water, even with a good wetsuit, any involuntary time spent under the water will not be a very pleasant experience. A cold-water hold-down will seem to last longer than a warm water one. When you do eventually surface, you feel a lot more drained of energy.

There are also a host of psychological factors that make cold waves seem heavier than warm ones. Warm waves are often an inviting light blue colour, conjuring up images of sunny days and fun on the beach. Cold waves are often an ominous dark green colour, conjuring up images of huge clean-up sets, strong rips and sharks. All this gets the adrenaline pumping before you even enter the water, let alone catch a wave or get a set on the head.

So, yes, waves are definitely ‘heavier’ in winter than summer, at least as perceived by us. It is the effect of the cold water on our bodies that without doubt overshadows the real increase in density and energy of a winter wave.

9 Local Winds on the Coast

Introduction

In this chapter we will talk about a phenomenon that may or may not affect us, depending on where and when we surf. For those lucky enough to surf where the climate is warm all year round, the sea breeze can be a daily occurrence. If you get up later than nine in the morning, chances are you will have missed the best surf of the day, as those light morning offshores start giving way to the inevitable onshores, making the waves mushy and of poor quality. If you make the effort to get up early, you will be rewarded by a couple of hours of uncrowded, glassy perfection, after which you can sit back and watch the crowds fighting it out over the midday onshore mush.

The **sea breeze** and its counterpart, the **land breeze**, are the daily onshore and offshore winds that occur on the coast, normally in warm climates or in summer. The phenomenon is a 'local' one, isolated to within the first couple of kilometres of the coast. It is directly attributable to the Sun's heating of the adjacent land and sea, throughout the day. In this chapter we will explain how it works and describe where and when its effect is most felt.

Recognizing the sea breeze

There are many different kinds of wind, each having its own particular characteristics. One example is the wind that blows around a low pressure – a wind produced by a large-scale pressure gradient, as described in Chapter 3. This is the essential ingredient for generating the waves we ride. Another is the **trade-wind**, felt on many offshore islands on the periphery of an anticyclone. The sea breeze is yet another, with its own unique set of qualities. It can be very useful to recognize which kind of wind you are dealing with, especially if you arrive at an unfamiliar surfing location.

Most places north of about 40° north do not have a sea breeze in the winter. So, if you arrive in Northern California one afternoon in

December and find the wind to be onshore, don't expect the conditions to be glassy the next morning. If you get the chance to look at a weather chart you will probably find a low-pressure system very near, with the isobars pointing straight towards the coast. In this situation, the best bet would be to look for a sheltered spot where those (typically) westerly gales might blow offshore.

In contrast, if you arrive at a tropical location mid-afternoon and the surf appears choppy, don't despair. It could be a sea breeze, or it could be the trade-winds. There is a good chance that, early the next morning, conditions will be glassy. Even if this spot suffers from cross or onshore trade-winds, quite often the land-breeze effect suppresses those trades in the early morning.

However, you may find yourself with a rare, six-foot early summer swell at a beachbreak in south-west France. It is a beautiful sunny day, and conditions are absolutely perfect at nine in the morning. Don't hesitate: get out there before it turns onshore.

Of course, there is great variability in the characteristics of the sea breeze. At some spots the surf will be blown out by ten in the morning, every day without fail. At others, you don't know when in the day the onshores will start blowing, if at all. Sometimes the wind will be offshore in the mornings, and sometimes there will be no wind. On rare occasions, the onshores might stop blowing mid-afternoon, or they might continue until well after dark.

A short-fetch windsea

One consolation with the sea breeze is that, when the wind drops, the waves clean up really quickly. This is very different from an onshore wind caused by a nearby low pressure, where the surf can take days to clean up, or might even drop off altogether before making a complete recovery. The reason for this is the length of the **fetch**, or the distance over which the wind blows. With a sea breeze, the fetch is extremely short, perhaps only a few kilometres. So, with a sea breeze, a few kilometres away from the coast there is no onshore wind at all.

When the breeze starts up, waves of very short wavelength are generated and degrade the surf. These short waves exist in addition to the waves you want to ride; surplus to requirements, if you like. If the sea breeze effect extends out to, say, 5 km offshore, these unwanted short waves will have been generated over a distance extending from 5 km offshore to the shore itself – a fetch length of 5 km (Figure 9.1).

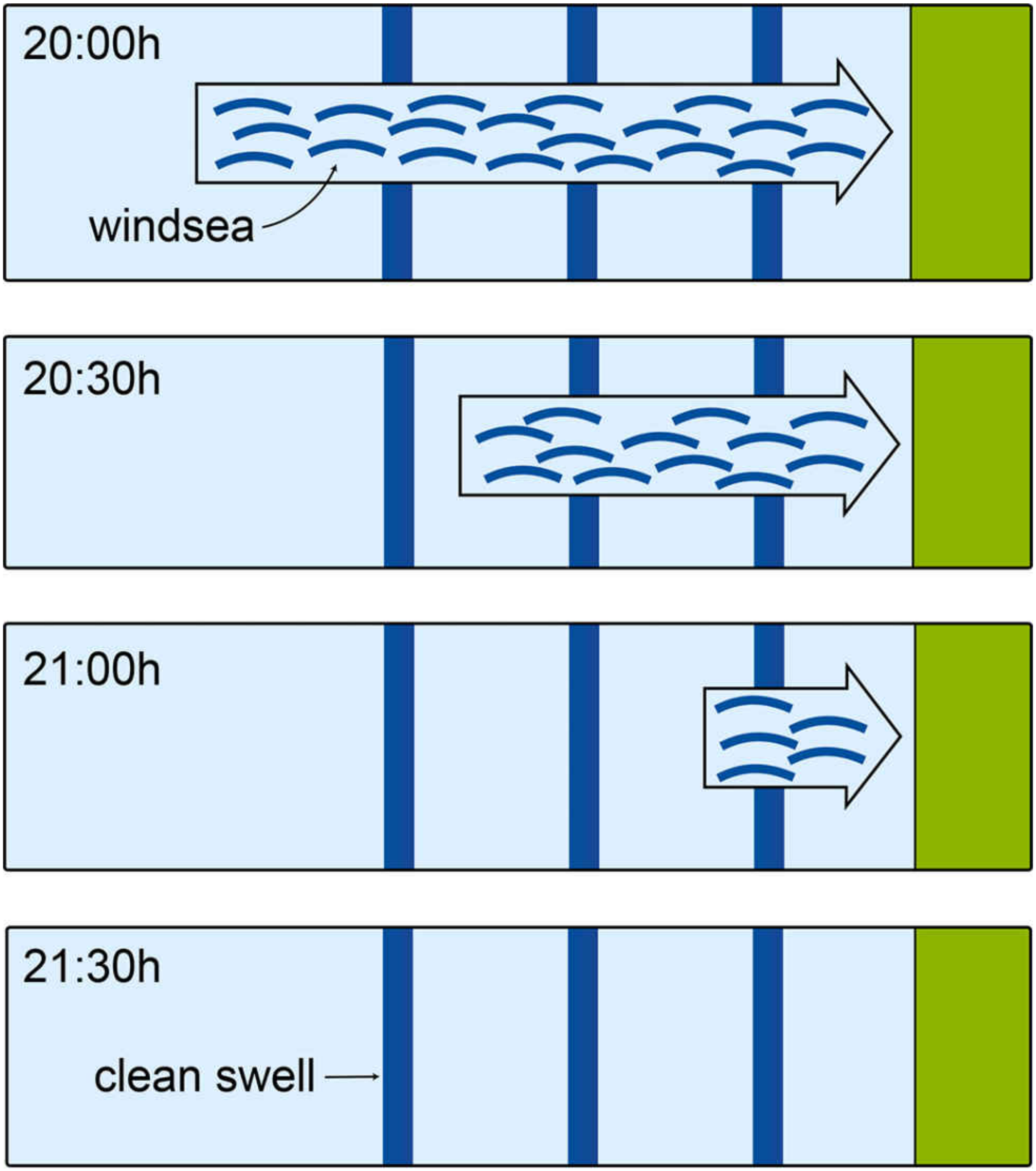

Figure 9.1: The sea breeze stops and no more short waves are generated; once they have all propagated to the coast, the surf is clean.

When the breeze stops, the short waves that were produced 5 km offshore still have to propagate to the coast. The time this takes depends on the propagation speed of the slowest waves, which are also the shortest, least-desirable ones. By the time the last waves reach the coast, the surf will have cleaned up. The short waves generated by the sea breeze no longer exist and you are left with nice, clean swell waves once again (Figure 9.1).

Compare this with a different kind of onshore wind where the fetch length extends maybe 100 km or more from the coast. Here, if the wind drops off, the last waves generated at the back of the fetch will have to travel at least 100 km before they reach the coast. In this situation, the surf will take a lot longer to clean up.

Note that, when the sea breeze starts blowing, its fetch extends only a little way off the coast. But as the day progresses and the strength of the breeze increases, the fetch extends further and further offshore until it reaches a maximum – which rarely exceeds about 15 km. Also, due to the Coriolis force (see Chapter 2), the sea breeze gradually turns around during the day, finally blowing at an oblique angle to the coast. In the Northern hemisphere, it turns to the right, and in the Southern hemisphere to the left, regardless of which direction the coast is facing. For example, on a west-facing coast in the Northern hemisphere, the sea breeze might typically start off blowing from the west. Then gradually, throughout the day, it veers, so that, by the time it reaches its peak in the afternoon, it is blowing from the north-west (Figure 9.2).

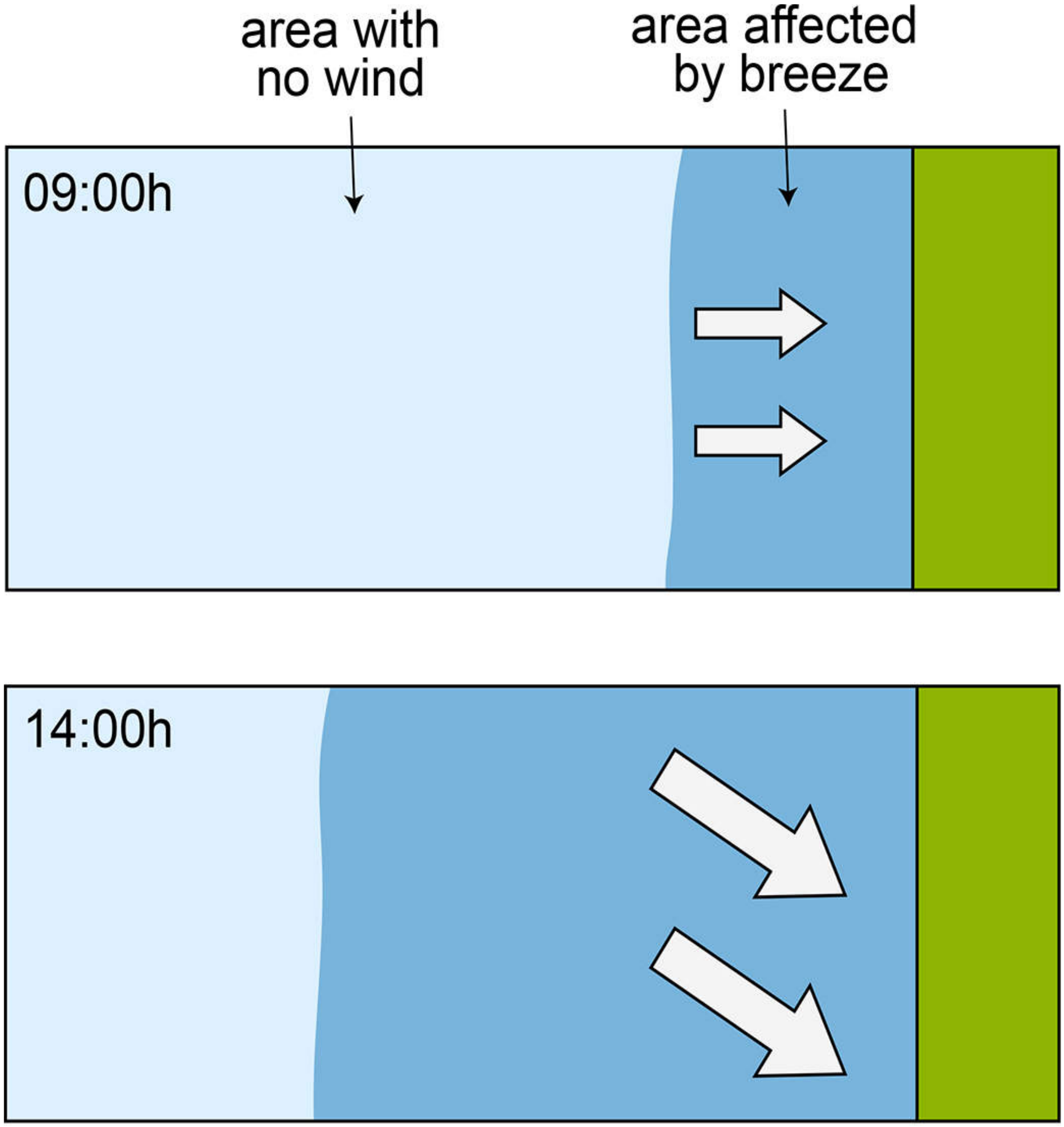

Figure 9.2: As the sea breeze develops throughout the day, its strength increases, the fetch increases and the angle changes. Northern hemisphere example shown.

How does the sea breeze work?

The mechanism responsible for the sea breeze is actually quite simple. It all stems from the fact that land and sea are two different substances, and they behave differently when they heat up and cool down. Land heats up and cools down very quickly, while the sea takes much longer to heat up and cool down. On a typical stretch of coastline, the sea might change its temperature only a few degrees throughout the day, while the temperature of the land might vary by tens of degrees.

This difference in daily temperature variation is due to the fact that water has a higher specific heat capacity (SHC) than land, as explained in Chapter 8. This means that, for the same amount of solar energy input, the land increases its temperature much more than the sea. The land is quick to respond to any energy input, unable to store much of that energy, so it raises its temperature very quickly. But the sea is sluggish, and responds more slowly, absorbing all that energy and increasing its temperature only very slowly.

In the morning, before the Sun gets up into the sky, either the land is slightly colder than the sea, or they are both around the same temperature. Throughout the morning, the Sun's energy heats up the land and the sea, but the land temperature increases much more than the sea temperature. So by mid-afternoon, the land is considerably warmer than the sea (Figure 9.3).

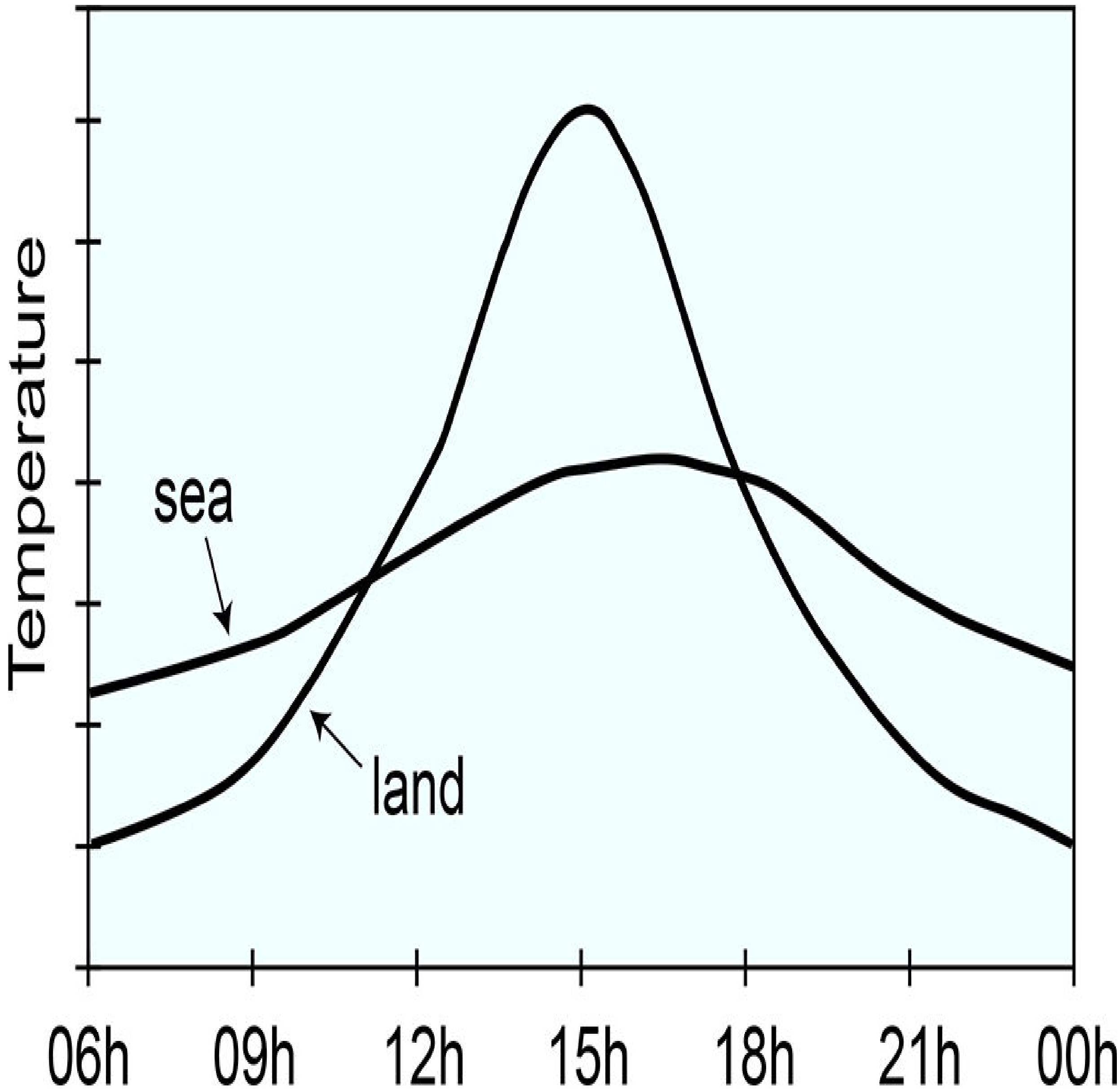

Figure 9.3: Typical variation of land and sea temperatures throughout the day.

The air over the land gets warmer, so it becomes less dense and starts to rise – a process called **convection**. The air overlying the sea does not rise because it is not so warm. The air that rises over the land leaves a gap, which must be filled, and the nearest thing available to fill that gap is the cooler air lying over the sea.

This filling of the gap generates a horizontal movement of surface air, called **advection**, from the sea to the land, leading to an onshore breeze, which continues like that for as long as the land remains warmer than the sea.

Later in the evening the land begins to cool, the sea breeze dies down, and you are left with glassy conditions. The air above the land has stopped rising, and may even begin to sink again. If the land temperature falls below that of the sea, the air drifts back from the land to the sea, producing a light offshore breeze. The land rarely gets much colder than the sea, so the land breeze is almost never as strong as the sea breeze. The land breeze will continue as long as the sea is warmer than the land, right on into the next morning if you are lucky (Figure 9.4).

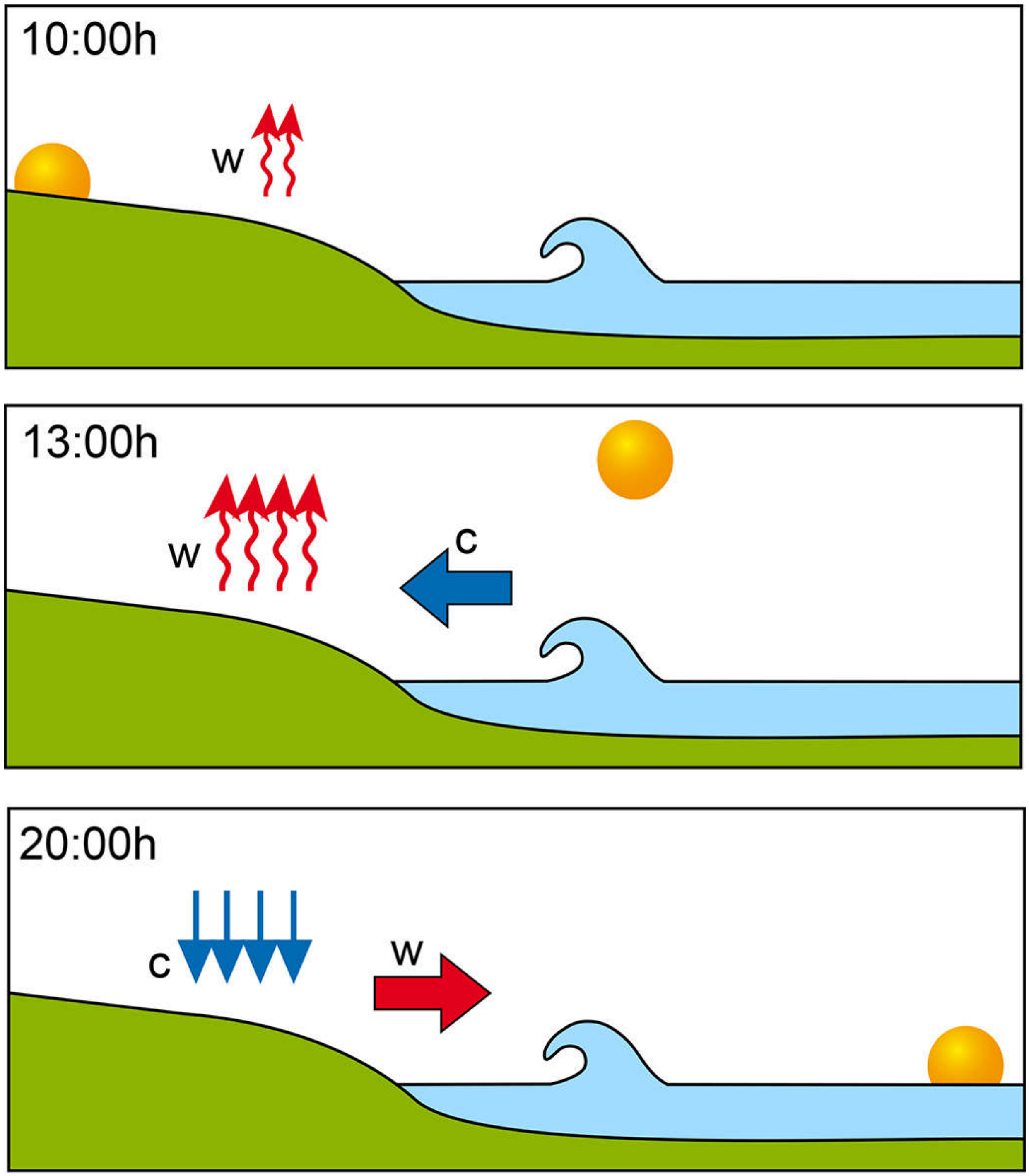

Figure 9.4: Typical daily sea-breeze cycle. The arrows marked 'c' are cold air and the arrows marked 'w' are warm air.

Where and when does sea breeze happen?

The principal requirement for a sea breeze is for the location to be generally hot and sunny. In latitudes where the climate is relatively cold, or it is cloudy most of the time, the sea breeze rarely occurs. The physical characteristics of the land and sea also affect the likelihood of a sea breeze. For example, if the land has something about it that enhances any temperature variations – makes it heat up and cool down more than it normally would – this will encourage the sea breeze.

The coastal deserts found on the western side of continents are places where conditions are perfectly suited to the sea breeze. The land has almost no vegetation, allowing it to heat up and cool down a lot, and extremely quickly. These places also tend to have cold surface currents, so the water is continually being replenished, and has little chance to change its temperature. In such locations, the land temperature might vary 20–30°C between night and day, while the water temperature remains practically constant. In these places, the sea breeze can sometimes become quite strong.

If you live somewhere where the sea breeze is strictly a summer occurrence, the characteristics of the breeze tend to change with the seasons. At the beginning of the summer, those nasty onshores will probably pester you almost every day. But as the months wear on, they get less and less frequent, until the autumn, when they more or less disappear altogether. At the same time, the offshores will tend to last longer into the morning. This is simply because the average sea temperature is getting progressively higher all through the summer, but the land temperature is still swinging up and down on a daily basis. So, during a sunny day at the beginning of summer the land tends to get much warmer than the sea, but at the end of the summer, it might get only a few degrees warmer. During the months when the air temperature gets well above the sea temperature during the day, you are likely to get a sea breeze, but in the months when the air temperature stays below that of the sea, a sea breeze is less likely (Figure 9.5).

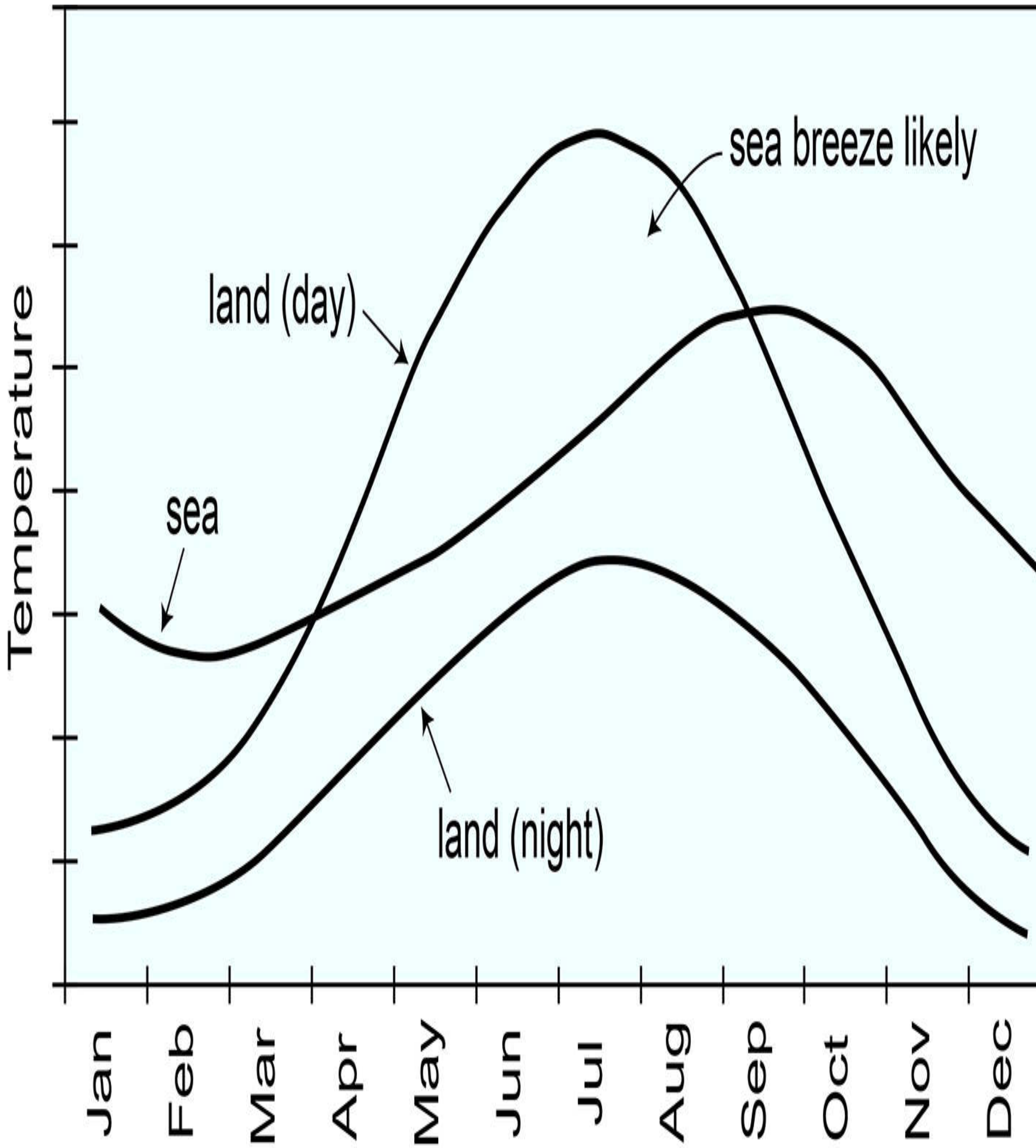

Figure 9.5: Typical land and sea temperatures throughout the year for some arbitrary place in the Northern hemisphere.

The sea breeze on an island

In theory, on an offshore island a sea breeze would develop all around the island, blowing inwards from the sea. The air would then converge and produce a giant column of rising air in the middle of the island. But in reality this doesn't happen. Islands are not perfect cones, and any variation in the topography around the coast tends to change the local characteristics of the sea breeze. In fact, the sea breeze might have radical variations from one spot to the next around the island.

In practice, if you are surfing on some mid-ocean volcanic island, especially one with high cliffs and valleys, and it looks like the onshores have set in, don't despair. The situation could be completely different at some other spot just around the corner. Here, you might find that the wind conditions remain calm for another few hours, or there may even be no sea breeze all day.

10 The Tides

This chapter might be a little bit more advanced than the rest of the book. Please feel free to skip it if you like. Of course, I'd prefer you read it.

Introduction

It almost goes without saying that tides are important not just for surfers, but for just about anyone who uses the coast. You might simply want to make sure you don't get cut off when walking the kids around the rocks at a seaside resort in south-west England; or you might want to be there at spring low tide in order to collect enough fish to feed your family for the rest of the month on a remote coral reef in south-west Madagascar. For us surfers, of course, the tides are vital. At most surf spots around the world, they affect wave quality in some way and, at a handful of those spots, they are the most important factor to deal with. If you are going on a surf trip, one of the most essential things to find out before you go is how the tides behave.

For many centuries, people have at least noticed the tides, even if they didn't know what they were or how they came about. Some early ideas now seem pretty bizarre. In biblical times, many people believed that tides were produced by the breathing of a giant whale; some thought the flood tide was caused by an angel sticking his foot in the water, and the ebb tide by him taking it out again. Around the Middle Ages, more advanced tidal theories started to bring the Moon into the picture, although not in quite the same way as we do nowadays. People had no idea of gravity, so they related the tides to the amount of moonlight falling on the Earth, which changed as the Moon altered its position in the sky throughout the month. One theory proposed that moonlight warmed the sea surface, releasing gases trapped in deep waters, which lifted up the surface of the ocean.

It wasn't until the seventeenth century that the basic principles of the workings of the tides, as we understand them today, were first laid out by Sir Isaac Newton. In 1687, Newton published his *Philosophiæ Naturalis Principia Mathematica*, perhaps the most fundamental scientific work of all time, and the world was made aware of gravity. Newton's theory – the **Equilibrium Tidal Theory** – was based on certain assumptions, many of which we now know were unrealistic; but it is nonetheless extremely useful for understanding the basics. Later on, the Equilibrium Theory was superseded by more complicated ones, such as the **Dynamic Tidal Theory**, introduced by Pierre-Simon Laplace in his monumental five-volume *Mécanique Céleste*, published between 1799 and 1825.

In the first part of this chapter, we'll look at the Equilibrium Theory, which deals with the driving forces behind tides, associated with gravitational forces and movements of the Earth, Moon and Sun. In the second part, we'll move on to the Dynamic Theory, which allows for certain factors that Newton was unable to consider, and, in doing so, brings us a little closer to finding out how the tides we observe behave the way they do. As in Chapter 2, where I described general atmospheric circulation, we will start with a very simple model, and gradually add more and more factors to make it more and more realistic.

1: The Equilibrium Tidal Theory

The Equilibrium Tidal Theory describes the way the height of the water on the Earth varies throughout the days, months and years, and how this is mainly due to the relative positions of the Earth, Moon and Sun. The gravitational force between the Moon and the Earth, combined with the gravitational force between the Sun and the Earth, is what generates our tides.

The Earth-Moon system

Let's forget about the Sun for a moment, and concentrate on the Earth and Moon together as a system. Most people think about the Moon revolving around the Earth, not the other way around. This is probably because the Moon is smaller than the Earth. If it were so small that its mass was insignificant compared with that of the Earth, the Earth would indeed just 'sit there' while the Moon revolved around it. The pivot point of the Earth-Moon system would be at the centre of the Earth. Now, imagine if the Earth and Moon were the same size. In this case they would both revolve around each other, with the pivot point located in space, half-way between the two bodies.

We can deduce that the position of the pivot point depends upon the relative masses of the two bodies: if one body is bigger, or heavier, than the other, the pivot point will always be nearer to the heavier one. In fact, instead of thinking about one body revolving around the other, it is better to think of them both revolving around their **common centre of mass**, or the natural balance point between the two. This is called the **barycentre**. In the case of the Earth and Moon, their relative masses dictate that the barycentre is actually inside the Earth, but quite near the periphery.

If we now think of the motion of each body separately, each one revolving around the barycentre, we can see that (a) the Moon describes a large circle because the barycentre is a long way away from the Moon, and (b) the Earth takes on more of a wobbling motion because the barycentre is actually inside the Earth itself (Figure 10.1).

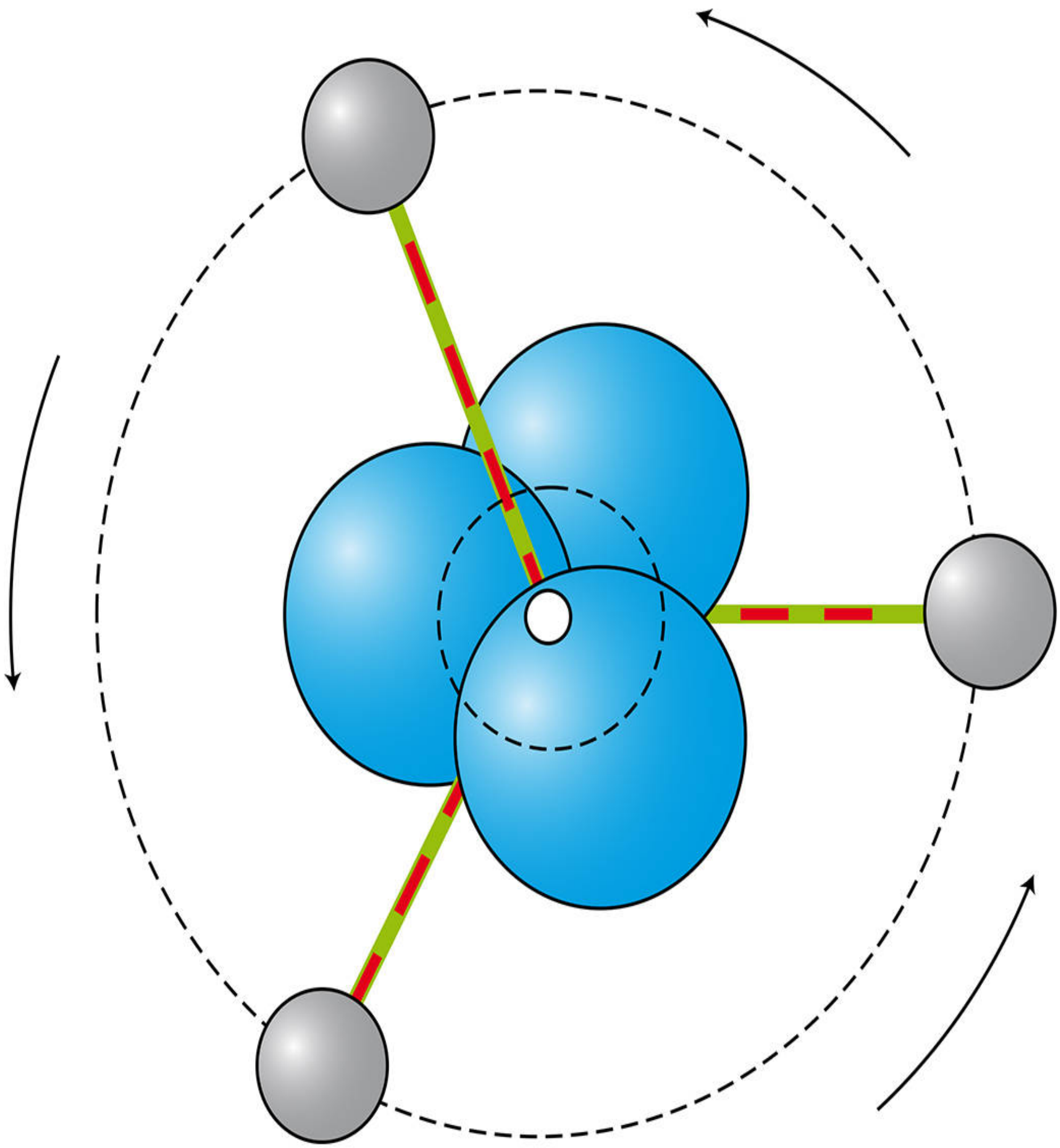

Figure 10.1: The Earth-Moon system revolves around the barycentre, shown by the white dot in the centre of the diagram.

If you can’t imagine the motion of the Earth very clearly, take a CD and stick a compass or other sharp object in it about 2 cm from the periphery; then swing the CD around on its pivot point. Don’t forget that the motion of the Earth around the barycentre has nothing to do with the Earth’s rotation around its own axis, giving us night and day. The two motions are completely separate. The Earth rotates around its own axis every 24 hours, whereas the Earth-Moon system revolves around the barycentre once every 27.3 days – called a **sidereal month**.

The Earth-Moon system is kept in equilibrium, circulating around the barycentre, by two forces. These forces are (a) the gravitational attraction between the two bodies, trying to pull them together, which we will call the **gravitational force**, and (b) the motion of the bodies themselves, trying to make them go off in a straight line, which we will call the **inertial force**.

Since there is no friction in space, the two bodies keep going like this. If you suddenly removed the gravitational force, the Earth and Moon would stop going around the barycentre, and, instead, they would fling off into space in a straight line. Likewise, if the two bodies didn’t have any initial motion, they would just crash together. We can think of gravitational force as a giant bungee holding the two bodies together. If the bungee breaks, they will fly off at a tangent, and, if something stopped them moving, they would be pulled together (Figure 10.2).

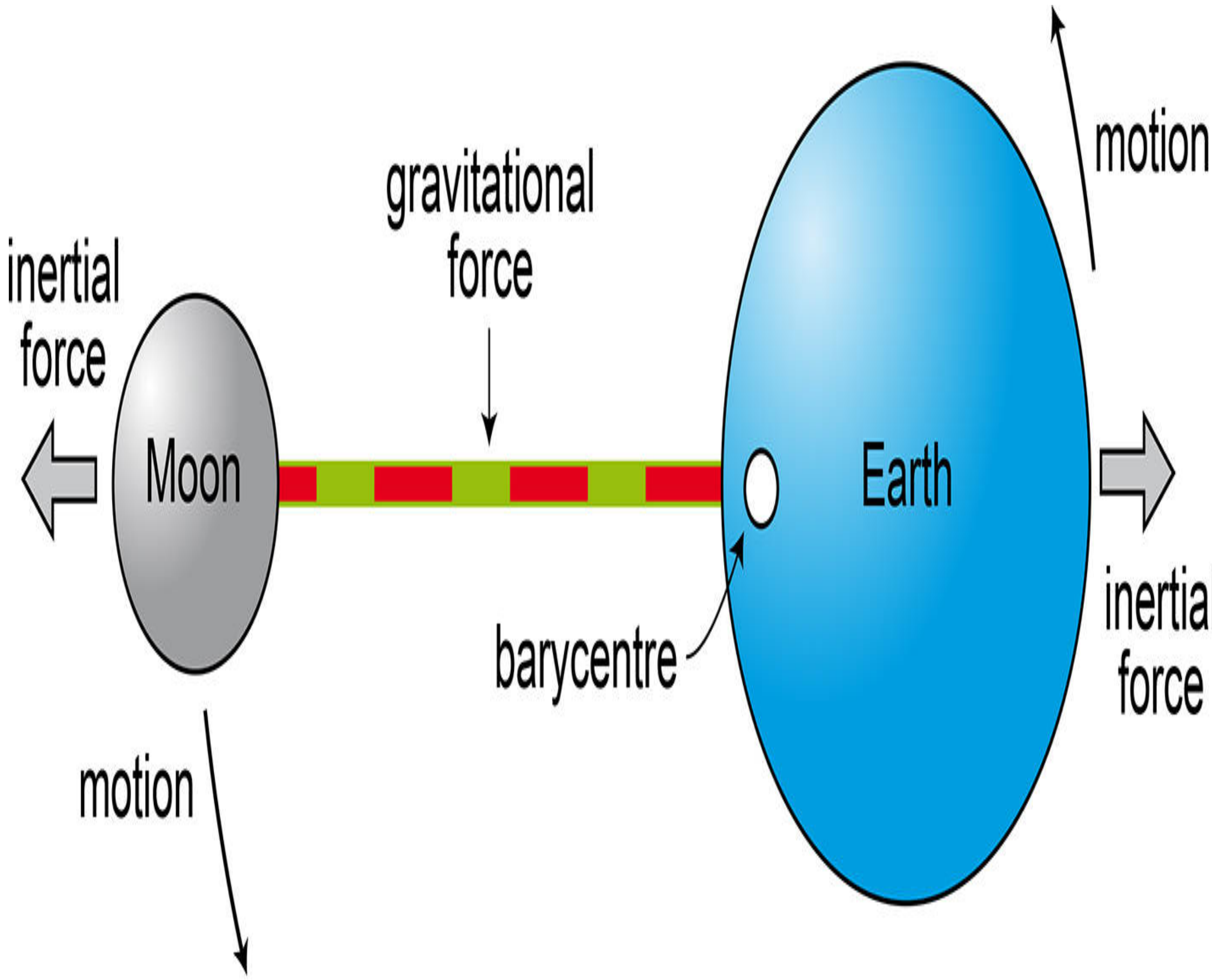

Figure 10.2: The forces keeping the Earth-Moon system in equilibrium are the gravitational force between the two and the inertial force due to their initial motion.

Tidal bulges

Now we come to the curious subject of **tidal bulges** – one that almost every textbook fails to describe properly. I'll try to do a little better.

Unless you live in Vietnam, Northern Australia or one of a handful of other places, you will experience approximately two tides a day. In Newton's model, where the Earth is covered in water, that water has two 'bulges' in it – one on the side facing towards the Moon, the other on the side facing away from the Moon. With these two bulges, it is fairly easy to see why there are two tides a day. The bulges stay in the same relative position aligned with the Moon, and the solid part of the Earth rotates on its own axis, every 24 hours, underneath them. Any point on the Equator passes underneath a bulge twice a day. An observer on the Equator would experience the bulges travelling from east to west like giant waves, with the passing of two crests and two troughs every day. So he would experience two high tides and two low tides a day (Figure 10.3).

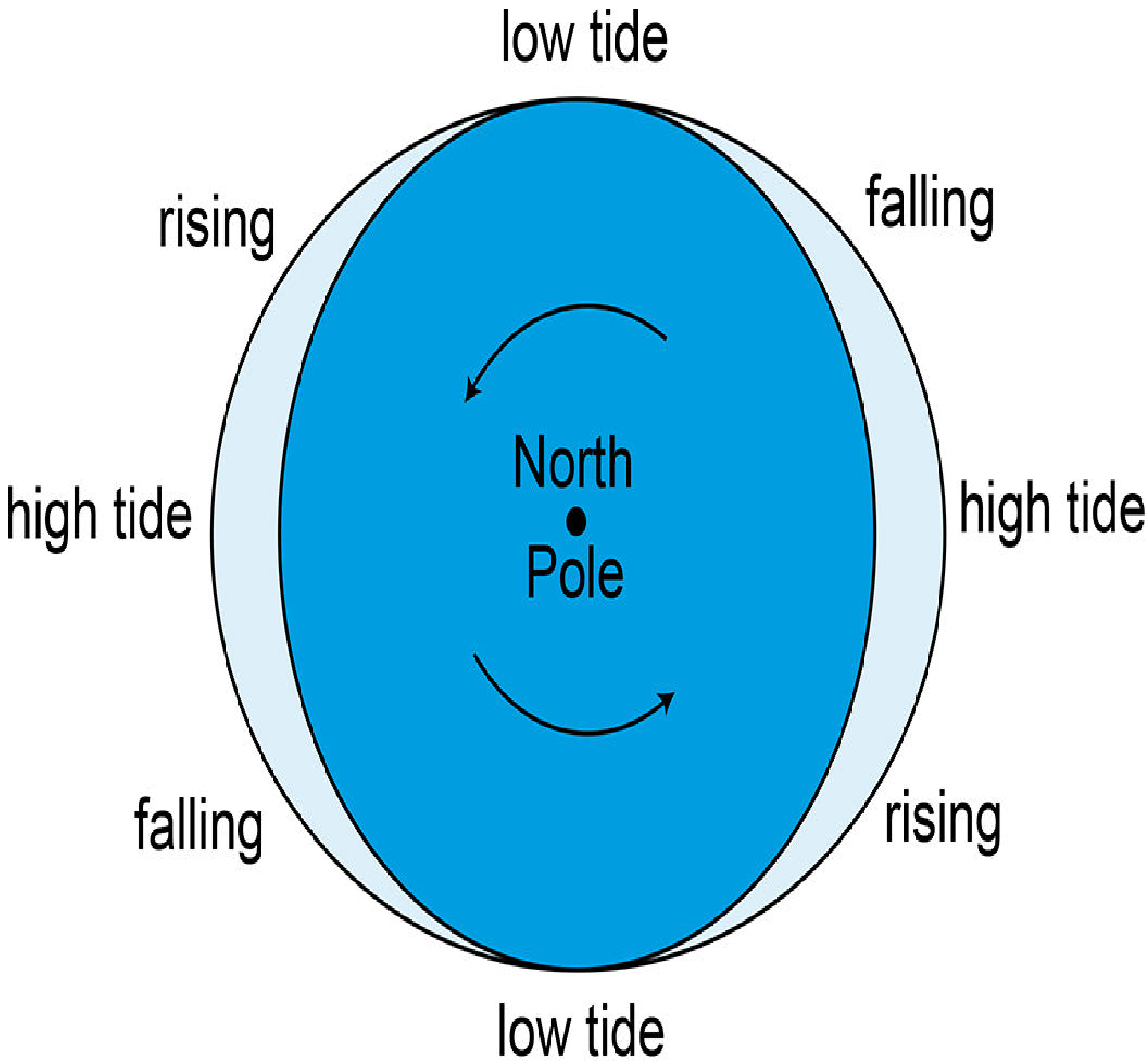

Figure 10.3: The Earth revolves about its own axis underneath the bulges, which, according to Equilibrium Theory, is what produces two tides a day.

But where do these tidal bulges come from? The Earth revolves around its own axis every 24 hours, which means that any given spot on Earth will experience the Moon passing through the sky every 24 hours. So, if the gravitational pull of the Moon has something to do with the tides, and the Moon passes over once a day, why isn't there just one tide a day? Why are there two tidal bulges, not just one?

Part of the answer to that question lies in the fact that it is not just the gravitational pull of the Moon that produces the tides. It is the gravitational force between the Earth and the Moon, combined with the effect of their relative motion. The tidal bulges are due to the balance between the two forces described above: the **gravitational force** and the **inertial force**. Here is the key: the balance between these two forces varies according to where you are on the Earth. The gravitational force is larger on the side of the Earth facing the Moon, and smaller on the side of the Earth facing away from the Moon. This is because the moonward side of the Earth is closer to the Moon, which means there is more gravitational force on that side than on the other side.

The inertial force is the same on both sides of the Earth. In fact, the inertial force *must* be the same all over the planet because the initial velocity that the Earth carries with it cannot be different for different parts of the planet.

When you combine the gravitational and inertial forces, you find that the **net force** – the resultant force when you add the two forces together – on one side of the Earth turns out differently from the net force on the other side. On the side facing the Moon, the gravitational force is bigger than the inertial force, so there is a net force towards the Moon. On the side facing away from the Moon, the gravitational force is smaller than the inertial force, so there is a net force away from the Moon.

As a result, the net force tries to stretch the moonward part of the Earth towards the Moon, and it tries to stretch the opposite part of the Earth away from the Moon. If the planet were flexible like a sponge, it would elongate itself towards and away from the Moon. In the Equilibrium Theory, the Earth is not very flexible, but it does contain a covering of water. So these two resultant forces pile up the water on opposite sides of the Earth, creating two bulges – one facing the Moon, and the other facing away from it (Figure 10.4).

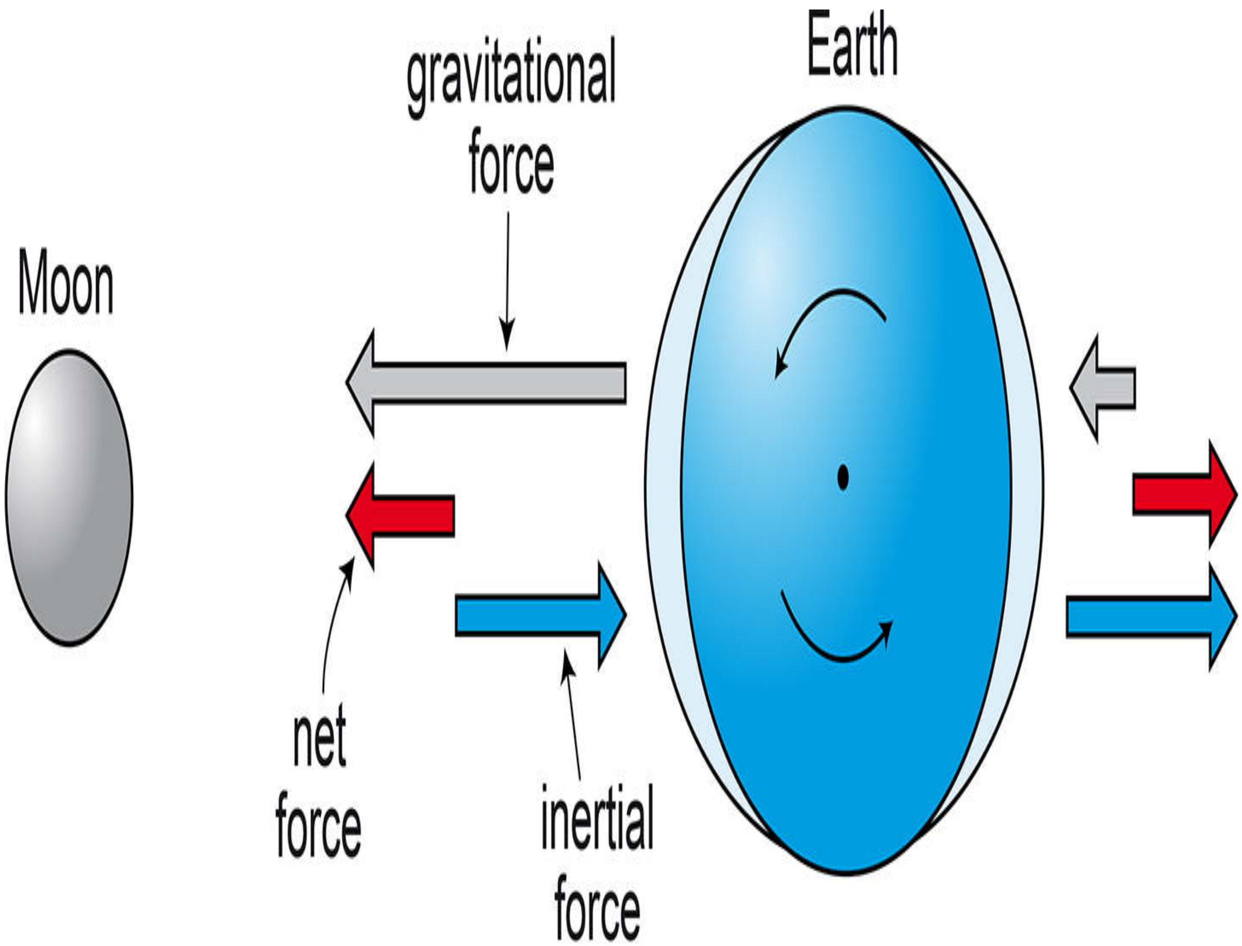

Figure 10.4: The gravitational force is different on each side of the Earth, whereas the inertial force is the same size both sides. As a result, the net force acts in opposite directions and produces the tidal bulges.

If you are still confused, look at the different arrows on Figure 10.4. On the moonward side of the Earth (the left), the gravitational arrow is bigger than the inertial arrow, so the net force is towards the Moon. On the opposite side (the right), the gravitational arrow is smaller than the inertial arrow, so the net force is away from the Moon. The only force that is a different size from one side of the Earth to the other is the gravitational force: the inertial force is the same size, and the net force is the same size but acts in opposite directions. Since it is the net force that produces the tidal bulges, the bulges turn out to be the same size on either side of the Earth.

The Earth-Moon-Sun system

You have probably noticed that the **tidal range** – the difference in the height of the water between high and low tide – is not always the same from one tide to the next. In the **spring-neap cycle**, the range varies throughout the month. The size of the tidal range grows and shrinks over about fifteen days, so we can say that the spring-neap cycle has a period of about fifteen days. So, for example, if at first the tidal range is at a maximum (high highs and low lows, i.e. **springs**), then just over a week later it will be at a minimum (low highs and high lows, i.e. **neaps**). And just over a week later than that, it will be at a maximum again, and so on.

The spring-neap cycle is associated with the phases of the Moon. Springs tend to occur near the time of full and new Moon, and neaps tend to occur near the time of half Moon. To understand why this is so, we have to bring the Sun into the picture.

First, it is useful to understand why we get full, new or half moons. This is all to do with the light from the Sun and the way it is reflected off the Moon.

> … When the Earth, Moon and Sun are lined up, and the Moon is in the middle, they are said to be in **conjunction**. If they were perfectly lined up, the Moon would pass directly in front of the Sun and there would be an eclipse. But most of the time this doesn't happen. Instead, a small proportion of the Sun's light glances off the Moon and reaches the Earth. What we see is a thin sliver of Moon – a new Moon (Figure 10.5, top panel).
> … When the Earth, Moon and Sun are lined up, and the Earth is in the middle, they are said to be in **opposition**. Again, most

of the time they are not perfectly lined up, so the Earth doesn't pass directly between the Sun and Moon and there isn't always an eclipse. Instead, the Sun's light is able to reach the Moon and bounce back directly to the Earth, giving us a full Moon (Figure 10.5, middle panel).

… When the Earth, Moon and Sun make up a right-angled triangle, they are said to be in **quadrature**. This is when we see only half the Moon illuminated by the Sun's light (Figure 10.5, bottom panel).

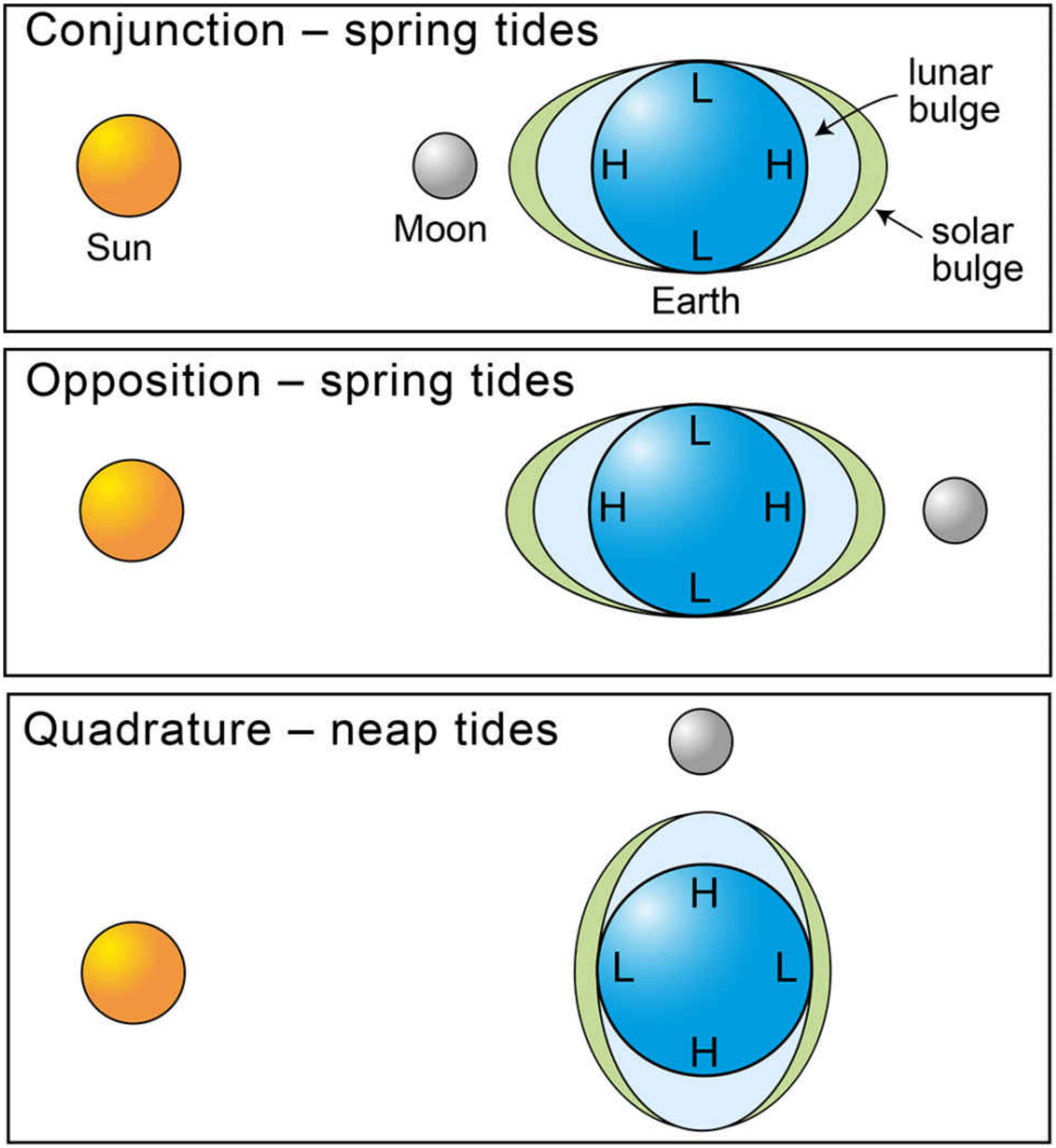

Figure 10.5: When the Moon and Sun are lined up you get spring tides, and when they are at right angles you get neaps. In the diagram, 'H' and 'L' mean high and low water respectively.

To understand how the spring-neap cycle is related to the different phases of the Moon, we need to think of the Earth-Sun system in the same way as the Earth-Moon system. Both bodies are rotating around a barycentre, and they are kept in equilibrium by the balance between the gravitational force and the inertial force due, as I explained earlier. This produces 'solar' tidal bulges as well as 'lunar' ones. The solar bulges are about half the size of the lunar ones, because the Sun's gravitational influence on the Earth is only about half that of the Moon.

It is the particular combination of the solar and lunar bulges – itself controlled by the relative positions of the Earth, Moon and Sun – that dictates whether the tides will be springs or neaps:

… ***Springs:*** When the Earth, Moon and Sun are in *conjunction* or *opposition*, the bulges are lined up, and therefore reinforce each other. This results in a large, combined bulge on two sides of the Earth and no bulge on the other two sides. There will be an extra-high tide where there is a combined bulge, and an extra-low tide where there is no bulge (Figure 10.5, top and middle panels).

… ***Neaps:*** When the Earth, Moon and Sun are in *quadrature*, the bulges are at right angles to each other, and so work against each other. This results in a lunar bulge on two sides of the Earth, and a solar bulge on the other two sides. There will be a lower-than-normal high tide where there is a lunar bulge, and a higher-than-normal low tide where there is a solar bulge (Figure 10.5, bottom panel).

Limitations of the Equilibrium Theory

Newton's Equilibrium Tidal Theory is a little oversimplified. It is fine for understanding the basics – the astronomical forces that drive the tides in the first place – but it is not really good enough to predict the tide with any useful accuracy at any given spot. The biggest problems are that (a) high and low tide at any part of the coast almost never occur at the times predicted by Equilibrium Theory, and neither do springs and neaps, and (b) the tidal range at some locations is typically much bigger than that predicted by Equilibrium Theory.

There are important reasons for this, mostly to do with the fact that Newton's model of the Earth is just not realistic. One serious oversimplification is that it assumes the planet is completely covered in water, with no continents. This is alright if we want to understand how the gravitational and inertial forces act on the planet to produce the tidal bulges, but that's about all. In reality, the continents seriously interfere with the movement of those bulges.

Newton was probably also unaware that the finite depth of the oceans has a significant effect on the tide. The tidal bulges are so large that their movement around the planet is governed by the depth of the water, making them slow down and refract over shallower areas such as the continental shelves.

Finally, there is the Coriolis force, which is not taken into account by the Equilibrium Theory. The Coriolis force makes objects or fluids turn to the right in the Northern hemisphere and to the left in the Southern hemisphere (see Chapter 2). The tides are influenced considerably by the Coriolis force, which diverts the water around imaginary axes called **amphidromic points**. Newton couldn't have known about this, because the Coriolis force wasn't discovered until over 100 years later.

2: The Dynamic Theory of Tides

In order to be able to explain and predict the tides in a practical way at different locations around the world, a completely different approach was needed. This didn't come about until Pierre-Simon Laplace introduced the **Dynamic Theory of Tides** around the early nineteenth century. This theory does not assume that the Earth is covered in water; it considers the effect of land masses on the tides. In addition, by thinking of the tidal bulges as giant waves circling around the planet, the theory is able to deal with both the finite depth of the oceans and the Coriolis force.

'Black-boxing' the Equilibrium Theory

Before we start to look beyond the Equilibrium Theory, we'll put away all that theory about the Earth, Moon and Sun in a 'black box', and instead just have a simple representation of the incoming forces that produce the tidal bulges. We'll call these the **tide-raising forces**. We can then go on to see what happens on the real Earth after the tide-raising forces are applied to the Earth from outside. A series of quite complex processes takes place between the 'arrival' of those forces and the resultant motions of the water on the planet.

We will think of the tide-raising forces as two giant magnets circling the Earth. They have to be special water-attracting magnets, otherwise they wouldn't work. The magnets complete one revolution of the Earth in just over 24 hours, and their rotation is in a clockwise sense looking down from the North Pole. Alternatively, you can think of the Earth rotating anticlockwise on its own axis underneath stationary magnets – the result is the same (Figure 10.6).

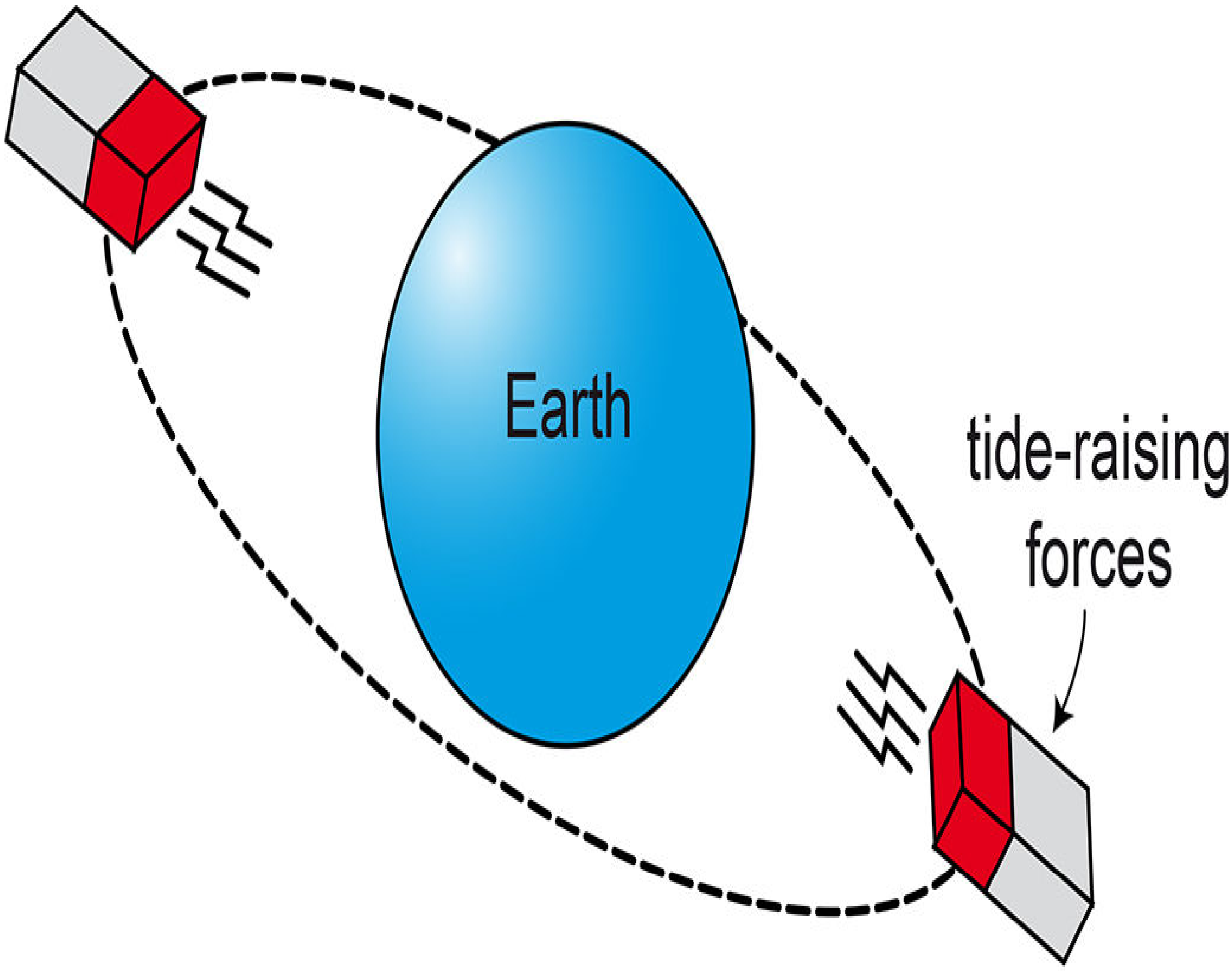

Figure 10.6: The tide-raising forces can be represented by two giant magnets circling the Earth.

The giant magnets are designed so that, on Newton's water-covered Earth, they would perfectly mimic the effect of the Earth-Moon-Sun system, producing two tidal bulges and two tides a day exactly as shown in Figure 10.3 and Figure 10.4. However, on a more realistic Earth, things are more complicated and the water can't respond exactly to the motion of the magnets:

> … First, the magnets try to produce tidal bulges that move across the oceans at the same speed as the magnets themselves, but the finite depth of the oceans slows down the bulges, so that they can't keep up with the movement of the magnets.
> … Then, when the bulges reach the edge of the continents, they can't simply propagate over the top of them – they bump into them and bounce back.
> … Finally, the bulges don't go in exactly the same direction as the magnets; they are deflected by the Coriolis force.

Tidal waves

To properly analyze those effects, a new approach to the tides was required. Instead of thinking in terms of simple tidal bulges, Laplace hit upon the key idea of considering the bulges as giant waves – **tidal waves** – circling the planet in a clockwise direction. By applying the rules of normal waves, he was able to analyze what happened as tidal waves propagated across the oceans and bumped into continents. Because tidal waves have incredibly long wavelengths (of the order of 10,000 km), they behave in special ways.

We'll start with the effects of finite depth. If the wavelength of a wave is more than about 20 times the depth of the water over which it is propagating, it can be considered a **shallow-water wave**. The speed of a shallow-water wave depends on nothing but the depth of the water (see Chapter 6). The deeper the water, the faster it goes. Normal, wind-driven waves are not shallow-water waves out in the open ocean because their wavelength (a few hundred metres) is usually much smaller than 20 times the water depth. But tidal waves are. Their wavelength is so long that they exceed the shallow-water wave criterion by a ridiculous margin. Even in the deepest waters of the world, the wavelength of a tidal wave is still around a thousand times greater than the depth of the water. So its speed depends only on the water depth.

According to simple wave theory, the speed of a shallow-water wave in km per hour is about 11 times the square root of the water depth in metres. In 4,000 metres of water – the average mid-ocean depth around the world – a shallow-water wave would travel at about 700 km per hour. The speed at which the tide-raising forces are trying to push the tidal waves around varies, but in many places it is more than 700 km per hour. As a result, the tidal waves lag behind and get out of sync with the tide-raising forces.

What is perhaps more important is that tidal waves travel at different speeds in different water depths. When they come out of the deep ocean and hit shallower water near coastal margins, they slow down. This makes them increase in height in a similar way to wind-driven waves. It is the basic idea of **shoaling**, and it is the reason why normal waves tend to 'jack up' before breaking (see Chapters 6 and 7). With a tidal wave, as the front of the 10,000-km long wave hits the shallow water of a continental shelf, it slows down, but the back of the wave is still in deep water, so it keeps travelling at its original speed. The wave is squashed up horizontally, and grows vertically. This amplification of the tidal wave over the continental shelf helps to explain why the tidal range – the difference in height between high and low water – in some coastal areas is much greater than predicted by the Equilibrium Theory.

Resonance

Now we'll add continents to the picture and see what happens to the tidal waves. Take the North Atlantic Ocean, for example. It has two almost north-south orientated boundaries: the Eastern Seaboard of North America and the west coasts of Europe. The tidal wave, which is trying to propagate from east to west across the Atlantic, doesn't simply pass over the top of North America; it hits the east coast and bounces off. A reflected wave then starts to propagate from west to east back across the Atlantic. When this wave gets to the other side, it once again bounces off the west coast of Europe, and so on.

This kind of wave, sloshing back and forth in a water body bounded at two opposite ends, is called a **standing wave**. You can easily create it on a much smaller scale using a rectangular washing up bowl. Just half-fill the bowl with water, slosh it from one end to the other and watch carefully. You will see that, when the water is high at one end of the bowl (high tide) it is low at the other end (low tide), and *vice-versa*. You will also notice that the overall vertical motion is at a maximum at the ends of the bowl, and zero at the centre.

If you had only given the bowl one quick slosh at the beginning, the standing wave would have quickly faded away to nothing, as friction took its toll and dissipated the wave energy. But if you gave it a small boost every time it came around to the same position, the wave would keep on going. By pumping a small amount of energy into the system at the right frequency, you compensate for the energy losses due to friction. This is exactly the same as pushing your child on a swing or your grandmother in a rocking chair; it is the main principle behind the internal combustion engine or any musical instrument. It is called **resonance**.

To keep a system like this going, the frequency with which you apply those boosts of energy must exactly match the **natural resonant frequency** of the system. For water bodies such as your bowl, this is the frequency with which the wave oscillates up and down the bowl. Put another way, the **natural period of oscillation** is the time taken for the wave to travel from one end of the bowl to the other and back again. It depends on the depth of the water (which governs the wave speed) and the length of the water body. Here are some examples of natural periods of oscillation:

… A bath, 1.5 m long, 20 cm deep: ***2 sec***.

… A swimming pool, 10 m long, 2 m deep: ***4.5 sec***.

… Loch Ness, 38 km long, 130 m deep: ***35 min***.

… Atlantic Ocean, 4,500 km long (‘wide’), 4 km deep: ***12.5 hrs***.

Where are we going with all this? We now know that, because of the shallowness of the water and the existence of the continents, the tide-raising forces can’t chase the tidal wave all the way around the planet like they did in Newton’s theory. But if they ‘catch’ the tidal wave at the right moment on its journey, they can give it a kick, as described above.

The last item on the list above states that the natural period of oscillation of the Atlantic Ocean is about 12.5 hours. The tidal wave is sloshing back and forth across the Atlantic about once every 12.5 hours. In Figure 10.5 you'll also see that the time taken for both magnets to circulate around the Earth is just over 24 hours – 24.8 hours to be more precise. So one magnet passes overhead every 12.4 hours, which is extremely close to the natural period of oscillation of the Atlantic. This means that the tide-raising forces (the giant magnets in Figure 10.6) are in sync with the Atlantic Ocean itself. Regular kicks from the tide-raising forces keep the water sloshing back and forth, like the water in your bowl or your grandmother in the rocking chair, but with period of just over 12 hours. At some spot, say on the east coast of the United States, there will be a high water once every 12 hours or so.

This effect doesn't only occur in the Atlantic – it is present in many other water bodies around the planet. As it happens, the majority of water bodies have their natural period of oscillation fairly close to 12 hours, so they experience **semi-diurnal** (twice-a-day) tides.

Amphidromes

There is just one more thing. I said that the tidal wave 'sloshes back and forth' across the Atlantic, and across the other oceans. Well, that's not quite true. Because of the Coriolis force, the motion is more of a 'swirling' than 'sloshing'. The fact that the Earth is spinning on its own axis has a deflecting effect on the motion of the water inside the tidal waves, which makes them swirl around imaginary axes called **amphidromic points**.

As explained in Chapter 2, the Coriolis force is an 'apparent' force, due to the spinning of the Earth about its own axis, that makes objects and fluids on large-scale trajectories veer to the right in the Northern hemisphere and to the left in the Southern hemisphere. The Coriolis force doesn't affect normal waves or open-ocean swells because waves don't actually 'carry water' from one place to another. There is no large-scale fluid motion, so there is nothing for the Coriolis force to deflect.

So if the tide is also a wave, why does the Coriolis force affect the tide? Tidal waves are so long that there *is* large-scale fluid motion. In a large ocean basin, the horizontal water motion beneath a tidal wave is large enough to be deflected by the Coriolis force.

Let's look again at the Atlantic Ocean. Instead of thinking of it as a simple water body with a boundary at each end, it is better at this stage to think of it as a large round bowl. Due to the Coriolis force, the water that is trying to travel horizontally back and forth is continually deflected to the right. Consequently, the water ends up hugging the periphery of the basin, and is forced to circulate around the bowl in an anticlockwise direction. It is like the standing wave described above, but with an added dimension. At the centre of the basin there is virtually no movement, and at the periphery is a wave that swirls around at the natural period of oscillation of the basin. The crest of the wave corresponds to high water, and the trough, on the opposite side, corresponds to low water. The tidal range varies from zero at the middle (the amphidromic point itself) to a maximum at the periphery.

If you are having difficulty visualizing this, replace your rectangular washing-up bowl with a large, round salad bowl. Instead of sloshing the water back and forth, swirl it around and around. The swirling water in the salad bowl closely simulates an amphidromic system. If you really want to see what's going on, video it and play it back in slow motion.

We are now starting to get a picture of the behaviour of real tides around the world. But to predict the exact times and heights of low and high tide at an exact spot on the coast, still requires quite a bit more theory. I won't go into many more details here, but will just say that the standard method used to predict tides is based closely on Laplace's Dynamic Theory of Tides. Instead of just the twice daily (semi-diurnal) tide described above, it takes into account lots of other cycles embedded within each other. Combining the tidal movement according to all these cycles gives a pretty accurate idea of tidal behaviour at thousands of spots along the coasts of the world.

Unpredictable things that can affect the tide

The tide tables produced for thousands of ports around the world are, in theory, highly accurate. They are based on the harmonic method of tidal prediction which, in turn, is based on the movements of Earth, Moon and Sun, as outlined in this chapter. Because it is based on the movements of celestial bodies, it is called the **astronomical tide**. However, ultimately, observed water levels at the coast don't depend on just the clockwork determinism of the astronomical tides; they also depend to a certain extent on local weather conditions. These, of course, are much more chaotic and unpredictable.

If, for example, there is a very strong onshore wind and large waves, they will literally push the water on to the shore – a phenomenon known to oceanographers as **set-up**. Gale-force onshore winds combined with a big swell can typically add up to 2 m to the height of high water.

Atmospheric pressure also affects the tides. Intuitively, you can think of the air pressure pushing down on the water surface – the higher the pressure, the more it pushes the water down. To be exact, for every millibar rise in atmospheric pressure, the level of the sea falls by one centimetre. Therefore, if you were in the middle of a 950-mb low pressure, the water level would be 1 m higher than if you had a 1,050-mb anticyclone sitting overhead. We can call this the **static pressure effect**.

If there is very low atmospheric pressure with a strong onshore wind and big swell (which often happens), these two effects combine to produce what is known as a **storm surge** – a phenomenon much feared by inhabitants of low-lying areas such as the Netherlands and south-east England. Considerable efforts have been made to guard against these storm surges, including the installation of the Thames Barrier – a kind of retractable dam designed to stop the whole of London getting flooded.

11 Rips

Introduction

Surfer 'A' is having a bad time. The waves are about 5 ft and hollow, with perfect rights and lefts, and the wind is a light offshore. But Surfer 'A' has been caught inside for ages, just duck-diving and duck-diving, seemingly forever. Every time he thinks he has broken through, another set breaks in front of him, pushing him back to the same spot. Meanwhile, just off to one side, Surfer 'B' coolly breezes by. With just a few lazy paddle strokes he's outside, without even getting his hair wet. Surfer 'A' is frustrated and bewildered; just before duck-diving the next wave, he sees Surfer 'B' coming out of the tube on the same wave.

Surfer 'B' is a bit smarter than Surfer 'A'. Instead of just blundering into the water, he took a few minutes to check the break from the top of the cliff. By seeing where the peaks were, how the waves were breaking, and where the rips were, he managed to save himself a great deal of time and energy. Where Surfer 'A' is stuck at the moment, the water is constantly flowing towards the beach and the waves are breaking with extra force. Where surfer 'B' is, however, the water is flowing out to sea and the breaking waves are either very weak or non-existent. Clearly, Surfer 'B' has the best strategy: use your brain to work around the situation rather than brute force to fight against it. While Surfer 'A' thinks of the rip as his worst enemy, Surfer 'B' thinks of it as his best friend.

For most surfers and people familiar with the coast, not knowing much about the rips can cause a lot of frustration and wasted energy; but for other people, it can be more serious. The typical once-yearly beachgoer from some inland city is usually reluctant to enter the sea where there are lines of whitewater rolling towards the beach (a natural mechanism for returning poor swimmers to the land), but he'll be keen to enter the water where the waves aren't breaking, oblivious to the seaward-flowing torrent ready to carry him well beyond the reach of those broken waves. 'Look,' he might say, 'those stupid lifeguards have put the flags up right where the waves are breaking. Let's go swimming over there instead, where it's safer.'

In this chapter we will look at the workings behind rips, and some of the different ways they can manifest themselves according to coastline topography and prevailing wave conditions.

How does a rip work?

Rips are nothing more than a mechanism for transporting water back out to sea again after it has been carried shoreward by the breaking waves. In deep water, way beyond the breakpoint, waves don't transport water anywhere (see Chapter 5). The particles just go round and round in circles and there is no net displacement of water. However, once waves have broken, there *is* a massive shoreward flux of water. The lines of whitewater, known by scientists as **bores**, actually carry water from the breakpoint to the shore. That's pretty obvious to anyone who has ever been in Surfer 'A''s situation. But obviously, all that water can't just keep piling up at the shoreline. Otherwise it would end up flooding inland, bursting its way through mountain ranges and over continents. So it must find its way out again. How does it do that?

That question is best answered by describing the most basic form of rip, on a typical stretch of beach with at least one or two sandbars along its length. Because the water depth is never exactly uniform along the shore, waves are focused on to areas of shallower water, and defocused away from areas of deeper water. This still happens even if the depth variations are very small. Once this process starts happening, it becomes self-reinforcing. The force of the waves, and hence the shoreward flux of water, is stronger where the waves are focused, and weaker where they are defocused. In the alongshore direction there is now an imbalance in the amount of water being pushed towards the shore. This, in turn, translates to an alongshore imbalance in pressure at the shoreline – higher pressure where there is strong onshore flux and lower pressure where there is weak onshore flux (Figure 11.1).

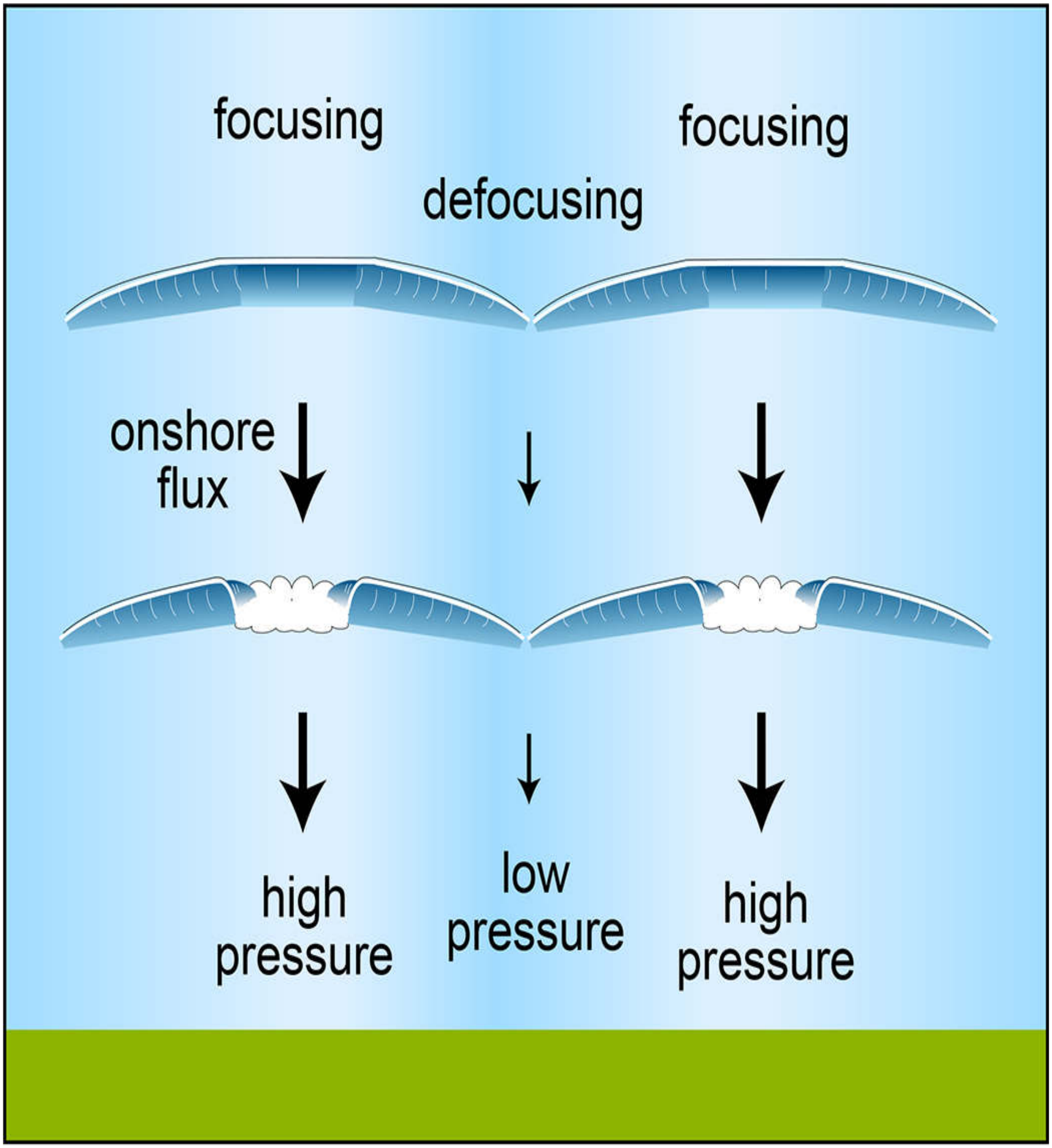

Figure 11.1: Alongshore pressure gradients at the shore caused by varying amounts of onshore flux of water.

Just like air flowing from higher to lower pressure in the atmosphere (see Chapter 2), the alongshore pressure-gradient causes the water to start to flow along the shore, diverging away from areas of high pressure and converging towards areas of low pressure. This alongshore flow is called the **feeder current**. As the feeder currents converge from either side, the water starts to force its way seaward, pushing against any weak onshore flux caused by the waves, flowing out past the breakpoint, and dissipating at the **rip head**. The water flows offshore at all depths, right down to the bed, so it drags the sediment with it and gouges out a deep channel. This, in turn, reinforces the focusing-defocusing mechanism we started with and maintains the alongshore pressure-gradients, keeping the system going (Figure 11.2).

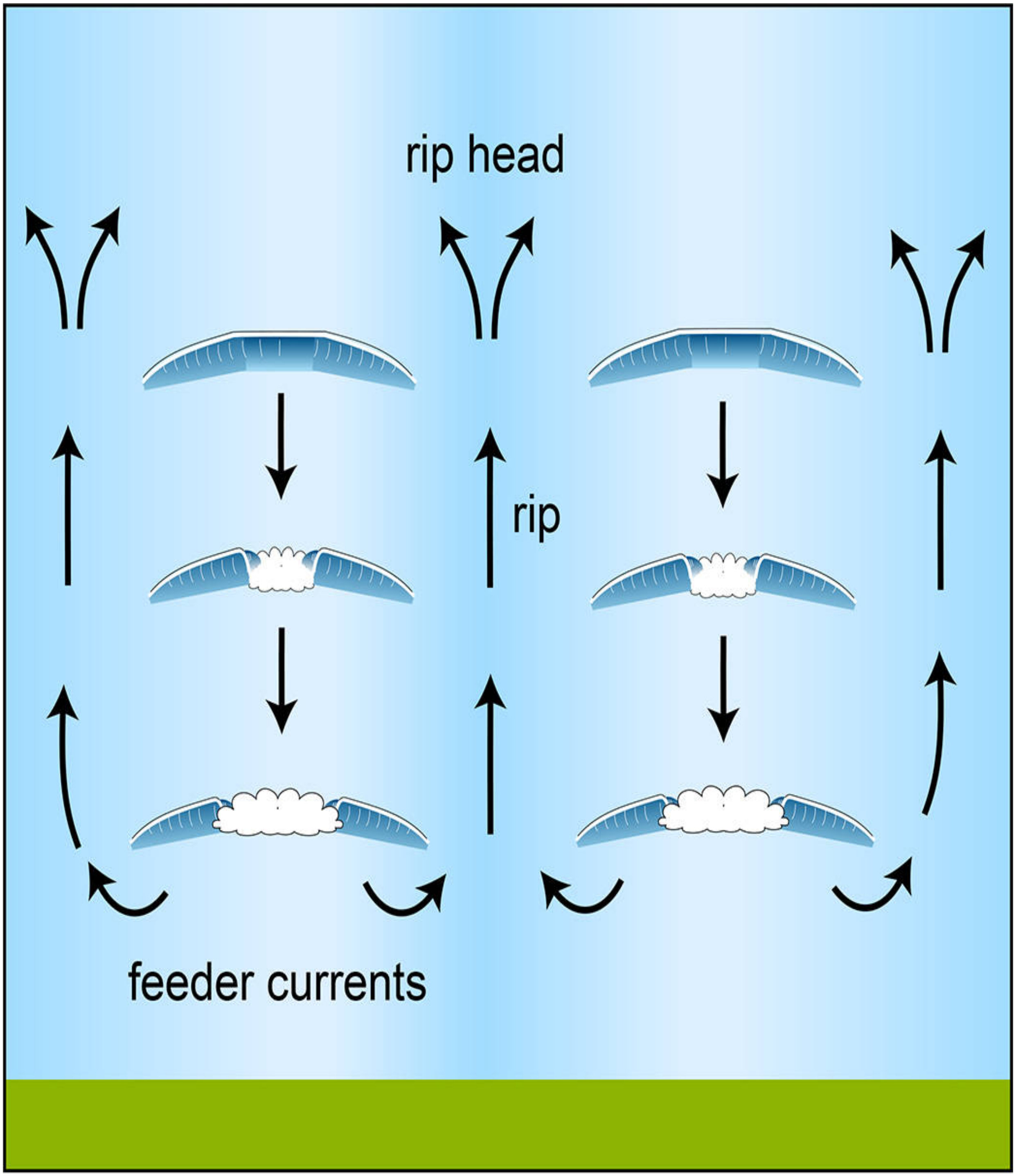

Figure 11.2: Fully-developed bar-channel rip system.

Types of rips

In order to makes things easier for future research, scientists always like to categorize natural phenomena according to their particular behavioural characteristics. Rips are no exception. If you are a surfer, it is also interesting to recognize the ‘official’ kinds of rips and how they compare with the rips we see when we are surfing. I’m going to describe some of the basic variations, and leave it to your imagination to derive more complex ones.

Accretionary beach rips are most similar to the basic beach rip described above. They normally occur when the wave height is less than about 1.5 m, and are associated with small-wave conditions when the sand is gradually moving onshore (the beach is **accreting**). Their spacing can vary enormously – from about 50 m to over 500 m – and they are able to stay in place for up to several weeks at a time. If the small-wave conditions continue for a long time, the sandbars gradually migrate shoreward and weld themselves to the shoreline, while the channels gradually fill in. If the wave direction is oblique, the whole pattern gradually migrates along the shore. As soon as there is a large storm or swell, the rip configuration is wiped out almost immediately and you start to get **erosionary** rips (see below).

Accretionary rips are by far the most common kind of rip, and statistically they are the ones that cause the most problems. This may seem strange, but it is simply because a much higher number of people enter the water during small waves, when the sea looks more inviting. The relative lack of danger is more than offset by the sheer number of people entering the water.

Erosionary beach rips occur when there are large or rising wave conditions, with wave heights in excess of 1.5 m. They are called erosionary because they are associated with coastal erosion, where the beach sediment gets carried offshore in the rip during high-wave conditions. What makes these rips special is that they are **hydraulically controlled**, not **bathymetrically controlled**, which means that their characteristics depend on prevailing wave conditions, not on the initial shape of the sea-bed. The incoming wave energy is so high that it controls the shape of the bed by quickly moving around large amounts of sediment. This, in turn, controls the position and size of the rips. In a growing storm or swell, the rips become larger, and get further apart, typically reaching several hundred metres between one rip and the next.

For inland tourists and poor swimmers, erosionary rips are statistically less dangerous than accretionary rips. This is because large waves and rolling lines of whitewater tend to put people off going in the water. The relatively high danger is offset by the much smaller number of victims.

Topographically-constrained rips are controlled by some physical feature sticking out above the water surface rather than the shape of the sea-floor itself. This could be a natural feature such as a cliff or headland, or a man-made one such as a pier or breakwater. The physical features of a system like this are likely to be fixed in position, so the rip is also usually fixed. At some spots, however, the rip can still vary quite a lot depending on the size and quality of the swell. Typically, you find a deep channel next to the topographic feature, and a shallow reef or sandbar off to one side. The waves break over the shallow area, and the rip runs out in the deep channel. The water carried shoreward turns alongshore in the feeder current, and is then deflected seawards by the topographic feature itself (Figure 11.3).

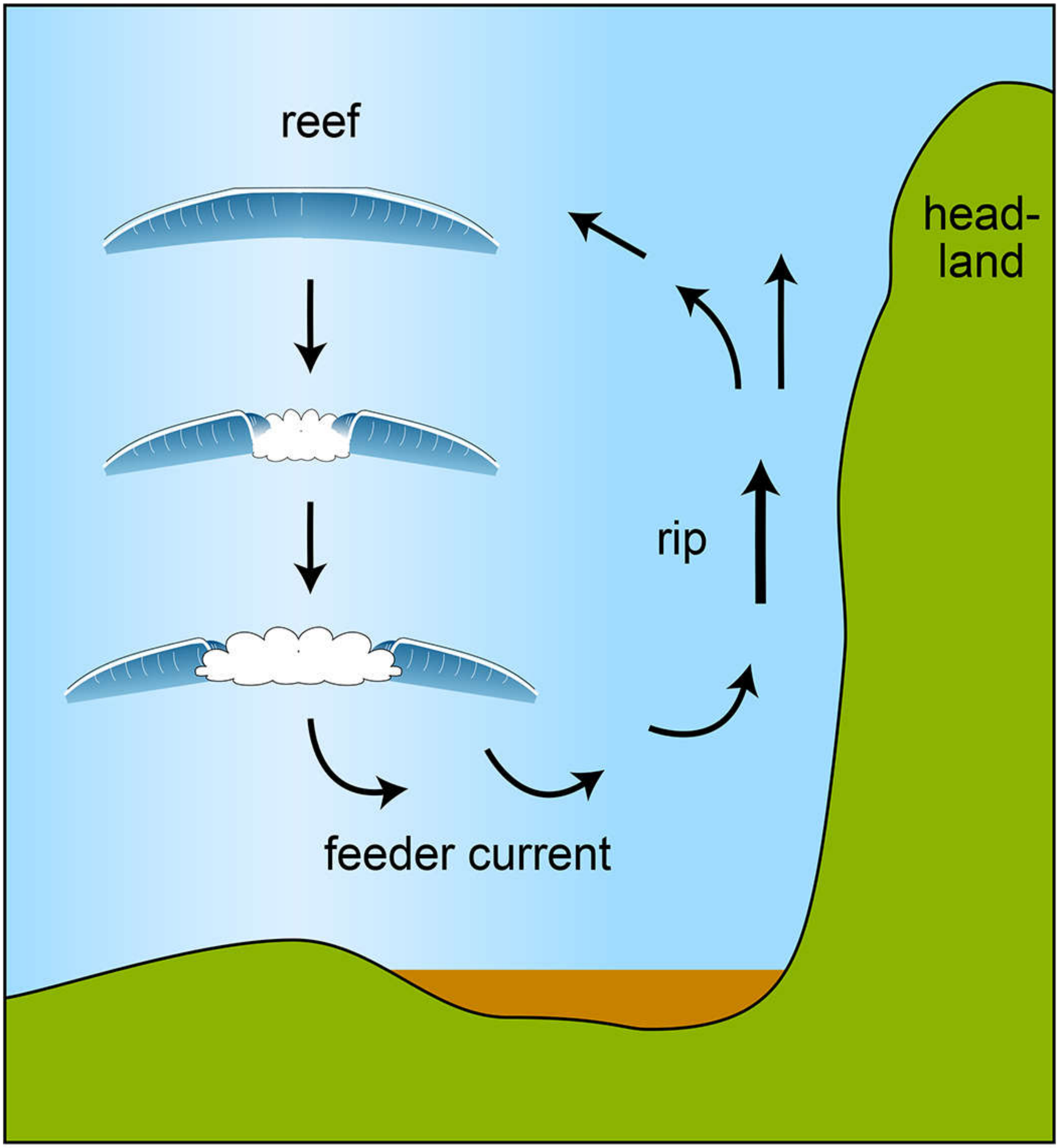

Figure 11.3: Topographically constrained rip. The arrows doubling back towards the peak (top right) show how this rip can also be rotational.

These rips are typically stronger than bathymetrically constrained ones, and the channel is usually deeper. This means that they pose a greater threat to potential victims. Often, the feeder current runs close to the shore between the shallow area and the topographic obstruction. Anybody entering the water here will get quickly swept away.

Rotational rips are rips that double back towards the breaking waves so much that they actually pass through the breaker zone itself. Instead of being dissipated at the rip head, the outgoing water is picked up at the breakpoint and carried shoreward again. The result is a closed circuit where the flow just goes round and round. Rotational rips are most common at big-wave reefbreaks where the rip is topographically constrained; but they have also been observed during smaller waves on beachbreaks. Experiments by my colleagues at the University of Plymouth, using floats with GPS receivers show that, in a rotational rip, the same water can flow around the circuit at least 15 times before dissipating.

Some big-wave spots are notorious for having rotational or at least semi-rotational rips on big days. The rip sometimes ends up right at the take-off zone, making the waves steeper, bumpier and more difficult to get into – additional factors you don't really need when it's big. Also, at these spots there might be only one way in to the beach. If you miss your opportunity and overshoot, and end up in the rip, the best option is usually just to relax, let the rip carry you back to the line-up, and then give it another go. Sometimes, especially if you've lost your board, it might take a few trips around to get it right.

Pulsating rips vary in intensity over timescales of around 5–20 minutes. They are what you get when the incoming waves come in distinct, well-defined sets. The speed and power of the rip constantly adjusts itself to compensate for the changing volume of water brought onshore by the waves. The effect is most noticeable at big-wave reefbreaks with fixed-position rips, but they can also occur on beachbreaks with smaller waves. Sometimes you can clearly see it happening: a short while after a set has passed, a large surge of water makes its way into the feeder current and then into the outgoing rip, dramatically increasing the speed and width of the rip itself.

On big days, a pulsating rip can be very dangerous for bathers. If most of the waves are breaking on a reef to one side of a deep bay, as in Figure 11.3, and the waves are coming in distinct sets with noticeable calm periods in between, the average beachgoer will probably be totally unaware of the size of the swell or the potential danger of the situation. The surge of water corresponding to each set might increase the water-depth near the shore by a metre or more. The beachgoer, having entered the water up to his waist, will suddenly get swept off his feet, into the feeder current and out to sea in the rip.

Megarips are simply defined as very large-scale rips that occur when wave heights are above about three metres. Usually, megarips are at least a kilometre long and have flow rates of three metres per second or more – about twice as fast as normal rips. When rips get this big, they start to act as extremely efficient sediment 'conduits', transporting huge amounts of sediment offshore and contributing majorly to coastal erosion. On long stretches of sandy coastline, the spacing between megarips gets progressively bigger as the wave height increases, sometimes reaching 500 m or more. If the coastline consists of adjacent bays and headlands, and the waves are breaking beyond the headlands, you might find one huge megarip serving two or more bays at the same time.

Alongshore drift is a kind of current produced by breaking waves, but not a rip in the true sense of the word. We'll include it here just because it is extremely common and highly relevant to surfing. The water moving in an alongshore drift comes partly from the onshore flux due to the broken waves, and partly from the feeder current. It occurs when the incoming wave direction is oblique to the shore, at a pointbreak for example. The water moves parallel to the shore in the same direction as the broken waves, and the strongest flow is usually between the breakpoint and the shore (Figure 11.4). The water is continually trying to escape out to sea, and may eventually do so if it finds an outlet point some distance down the coast; or it may continue running along the shore for many kilometres. Alongshore drift can be annoyingly strong. At some very long pointbreaks, the current is so strong that getting out of the water and walking around is much more practical than paddling back to the take-off spot.

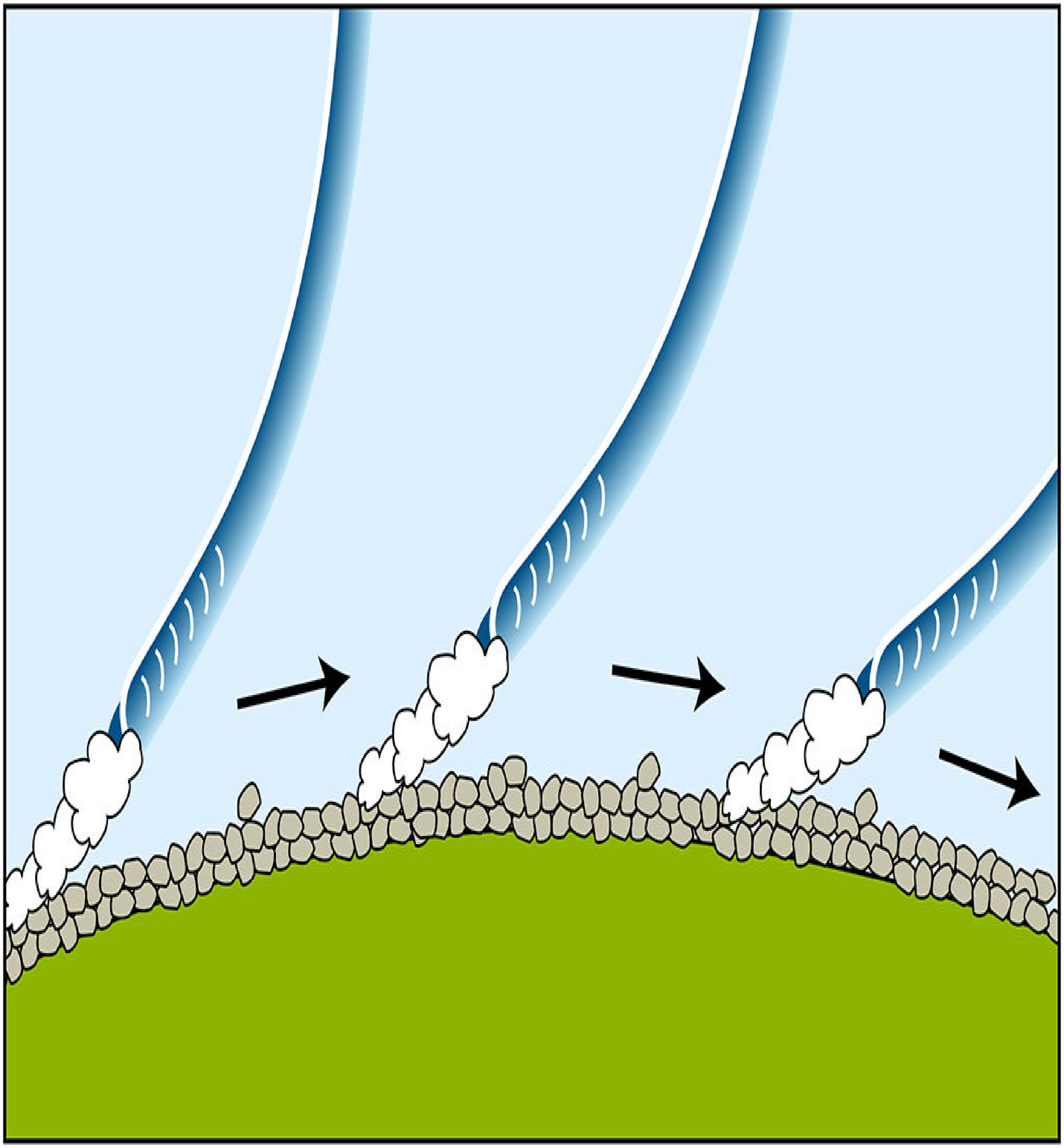

Figure 11.4: Alongshore drift.

Just like any natural phenomenon, most rips don't exactly fit any one of the above categories, but by combining the characteristics of two or more basic kinds, you can probably describe in great detail the rip at your local spot on any particular day. Sometimes though, rips defy definition altogether, and seem to have a mind of their own, just like the waves that caused them.

12 Forecasting 1: Basics

Introduction

Until quite recently, surf forecasting was shrouded in a certain amount of mystique – almost a ‘dark’ science, best left to a handful of gurus endowed with some special power. For most of us, wave forecasting was little more than blind guesswork. Nowadays, any of us can access a swell-forecasting website that will tell us where to go, what board to take, what wetsuit to wear, and what stickers to put on the board; no brain required. The latest surf-forecasting websites provide site-specific values of a single height, period and direction, plus various ‘star-rating’ schemes, seemingly drawn from a black box that we, the punters, cannot see inside.

Most of the time, these kinds of surf forecasts are fine. They give us more than enough details to decide which beach to head for, which board to take, or whether to go to work tomorrow. But sometimes, when ‘they’ say it’s going to be six foot and offshore, and it isn’t, we wonder why. What we perhaps don’t realize is that, sometimes, the information we are given is not complete enough to describe exactly what the waves will be like when they reach the shore. This is when we need to look at the bigger picture. Where exactly did that swell come from; what trajectory did the storm that produced it take; what were the winds like, say, 400 km from the coast a few hours before the swell hit… and so on.

To do this, we need to back-track a little. We must resort to some of the ‘old-school’ wave-prediction tools, the most useful of which is the isobar chart (see Chapter 3). Also very useful are wind charts, wave-height and period charts, and wave buoy measurements. When I do my own wave predictions, I never just look at site-specific forecasts; I check a whole range of charts and measurements at the same time, and the most basic isobar charts are of utmost priority.

In the first part of this chapter, we’ll look at some of these basic resources; in the second part, we’ll talk a little about wave predictions and their reliability.

Don't forget, internet surf-forecasting is evolving extremely rapidly. I've tried to make this chapter as generic and 'timeless' as possible, but, inevitably, by the time you read this, something new will have appeared.

The original surf-forecasting tool: the isobar chart

By now, you will have a good idea of what isobar charts are and what they do. To summarize, isobar charts, or **surface pressure charts**, show the atmospheric pressure field over a designated area of the planet, at a height just above sea level, normally about 10 m. Atmospheric pressure is measured in hectopascals (hPa) or millibars (mb), which are the same thing. The weight of the entire atmosphere on the ground exerts a pressure of about 1,000 mb; but depending on the conditions, this may vary from about 900 to 1,050 mb.

The charts depict the pressure 'field' by using contour lines of equal pressure, called isobars (*iso* = equal, *bar* = pressure). The closer the isobars, the stronger the pressure gradient – the change in pressure over a certain distance. A stronger pressure gradient means a stronger wind; therefore, closely packed isobars mean strong surface winds which, in turn, lead to bigger surf. The wind blows more or less along the isobars, but sometimes up to about 15° pointing inwards towards the centre of a low pressure. In the Northern hemisphere, the wind blows anticlockwise around a low pressure and clockwise around a high pressure. In the Southern hemisphere, this is reversed.

The surface pressure charts give us, through the pressure field, an idea of wind direction and strength over a particular area of the planet. They are also good graphical illustrations of key features such as low pressures. Some charts also show the position of **warm fronts** and **cold fronts**. A warm front has relatively cold air in front of it, and warm air behind it; the reverse is true of a cold front. These fronts also mark a sudden shift in wind direction, which is useful to know, because the passing of a front can either clean up or destroy the surf in very little time.

In the early days, we had to rely on a surface-pressure **analysis chart** published in a newspaper or shown on television. For a long time, this was the only resource available to us for forecasting surf. An analysis chart is compiled from a series of atmospheric pressure values measured by a network of buoys and ships, which gives a 'snapshot' of the pressure field at some time in the recent past (Figure 12.1).

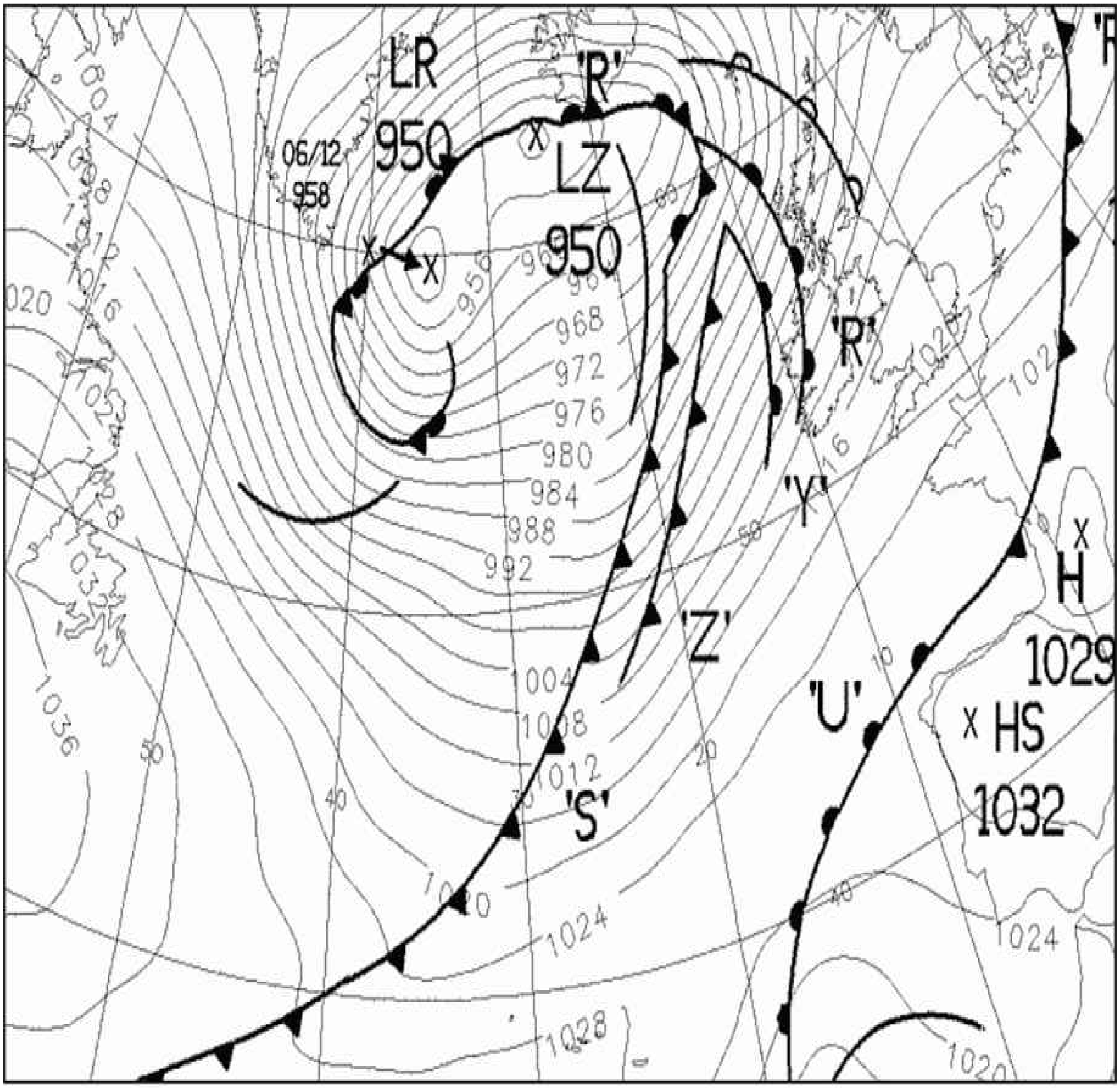

Figure 12.1: North Altlantic isobar chart. Numbers on the isobars show pressure in mb; thick lines with round blobs are warm fronts; those with triangles are cold fronts.

From this, we tried to make a reasonable estimate of surf conditions in the near future. From the orientation and number of isobars we could identify the current position of a storm over the ocean, together with its all-important fetch (the area containing those winds that generate the swell we would eventually ride, see Chapter 3). With some experience of the likely size of swell produced by a particular fetch, how fast it would travel and how it might be affected by the continental shelf, we could make a guess at its height and arrival time. And with some previous knowledge of the most likely trajectory of storms in that particular ocean, at that time of year, we could make a guess at local wind conditions expected on that beach.

That's a lot of guesswork based on not much information. Even for the most experienced and talented wave-forecasting gurus, it was still very difficult to predict the state of affairs at the end of a long and uncertain string of processes connected to just one snapshot of a single parameter. There were just too many variables.

After a while, special television forecasts began to give **forecast charts** for the next few days ahead. You could also get them in the form of a fax, usually by subscribing to a special marine service. Forecast charts give much more than analysis charts: they show us the expected trajectory of the storm, its persistence over the ocean and the local wind conditions for when a swell is about to arrive. They are based on simulations of the atmosphere in a computer model, not just on the measurements from buoys and ships.

Wave prediction charts

Wave prediction charts give an analysis and forecast of wave characteristics over a designated part of the ocean. They are very useful if you want to see the predicted evolution of a swell over the ocean during the next few days, and to track its journey from the storm centre to the coast. By opening up the field of view, this can add a bit more clarity than just having data for a specific point on the coast, as in most surf-forecasting websites. For example, the wave height at your local beach might jump from zero to two metres next Monday, but you wouldn't know where that swell came from if you only had a site-specific forecast.

The main parameter of interest is wave height, and this is normally indicated on the charts by different coloured contours representing different ranges in wave height. The normal convention is to have ‘cold’ colours at the low end of the scale, and ‘hot’ colours at the high end. Other parameters may also be shown, depending on who produces the charts. For example, you might see wave period, direction of wave propagation, wind strength and direction (Figure 12.2).

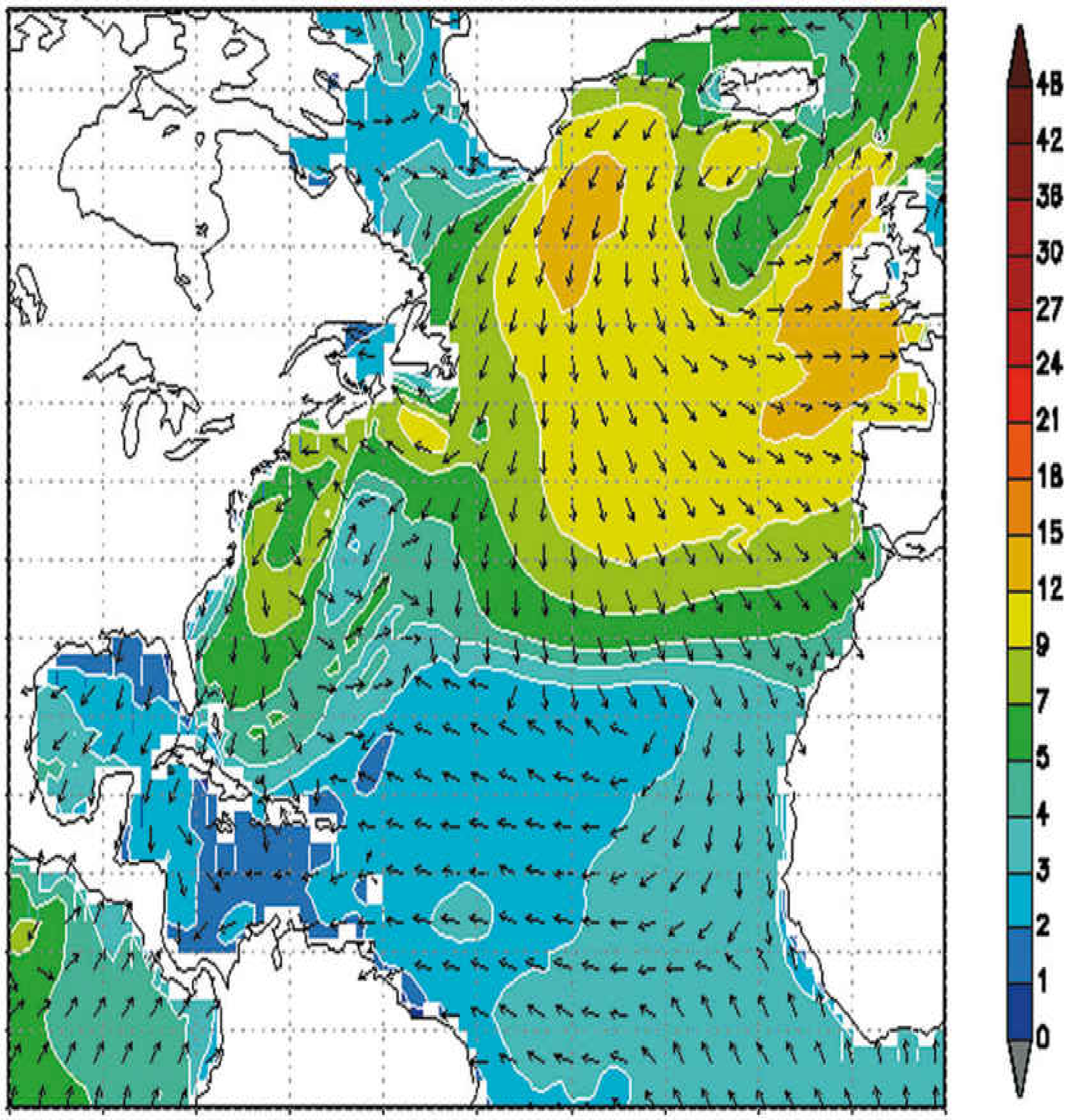

Figure 12.2: Typical wave prediction chart showing wave height in feet (shaded contours) and propagation direction (arrows).

Wave period (usually **peak period**, which I will explain below and talk more about in Chapter 13) is shown on these charts with colour contours, the ‘cold’ colours corresponding to short periods and the ‘hot’ colours corresponding to long periods. In general, although not always, longer periods mean cleaner swells. For example, a period of sixteen seconds would probably indicate a new, clean swell from a distant storm, whereas a period of six seconds would most likely indicate a poor-quality windsea. On some wave period charts there is a thick, dark line called the **swell front**, which represents a region of abrupt change in period. Since the waves at the beginning of a swell have a longer period than those at the end (radial dispersion – see Chapter 5), the swell front, marking a sudden change from short to long periods, can be used to work out the arrival time of a new swell. This is extremely useful to us. As soon as the dark line hits the coast, you know that the first waves in the swell will be arriving (Figure 12.3).

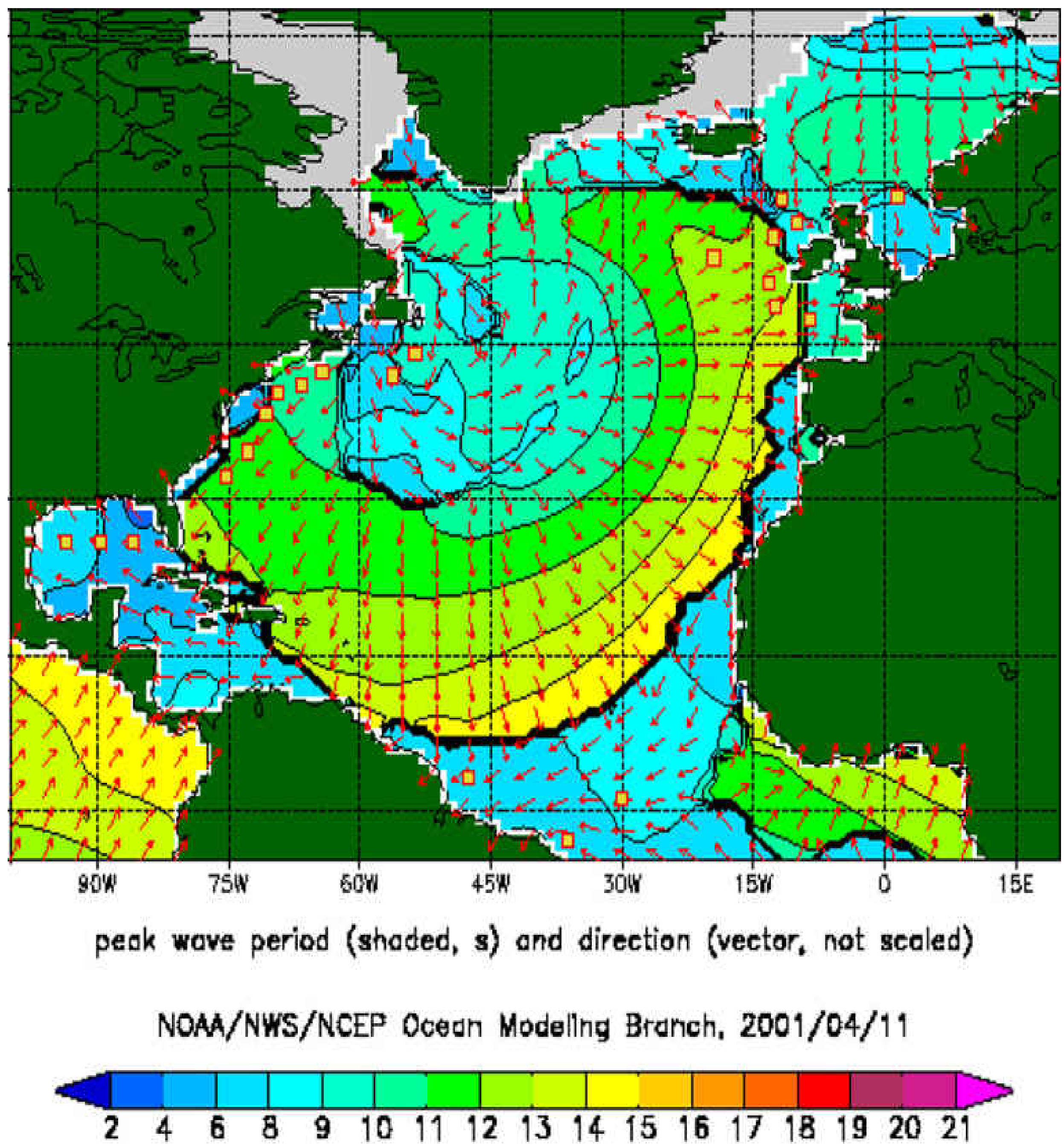

Figure 12.3: Typical chart showing wave period in seconds (shaded contours) and direction (arrows).

Wind prediction charts

Wind prediction charts work on the same principle as wave prediction charts, only instead of wave height, period and direction they show predictions of wind strength and direction (the 'windfield') over the ocean. Wind charts complement isobar charts very well. While isobar charts are very useful for identifying position and tracking movement of low- and high-pressure systems, wind charts enable us to quickly identify strong fetches and track their evolution over time.

Wind strength is depicted by colour contours, with 'cold' colours meaning light winds and 'hot' colours strong winds. Wind direction is sometimes shown by arrows, and sometimes by barb symbols (see below). Note that the arrows point in the direction of 'propagation' of the wind, even though wind direction is always quoted as the direction from which it comes. For example, an arrow pointing towards the north means a south wind (the wind is coming from the south, but the air is 'propagating' towards the north) (Figure 12.4).

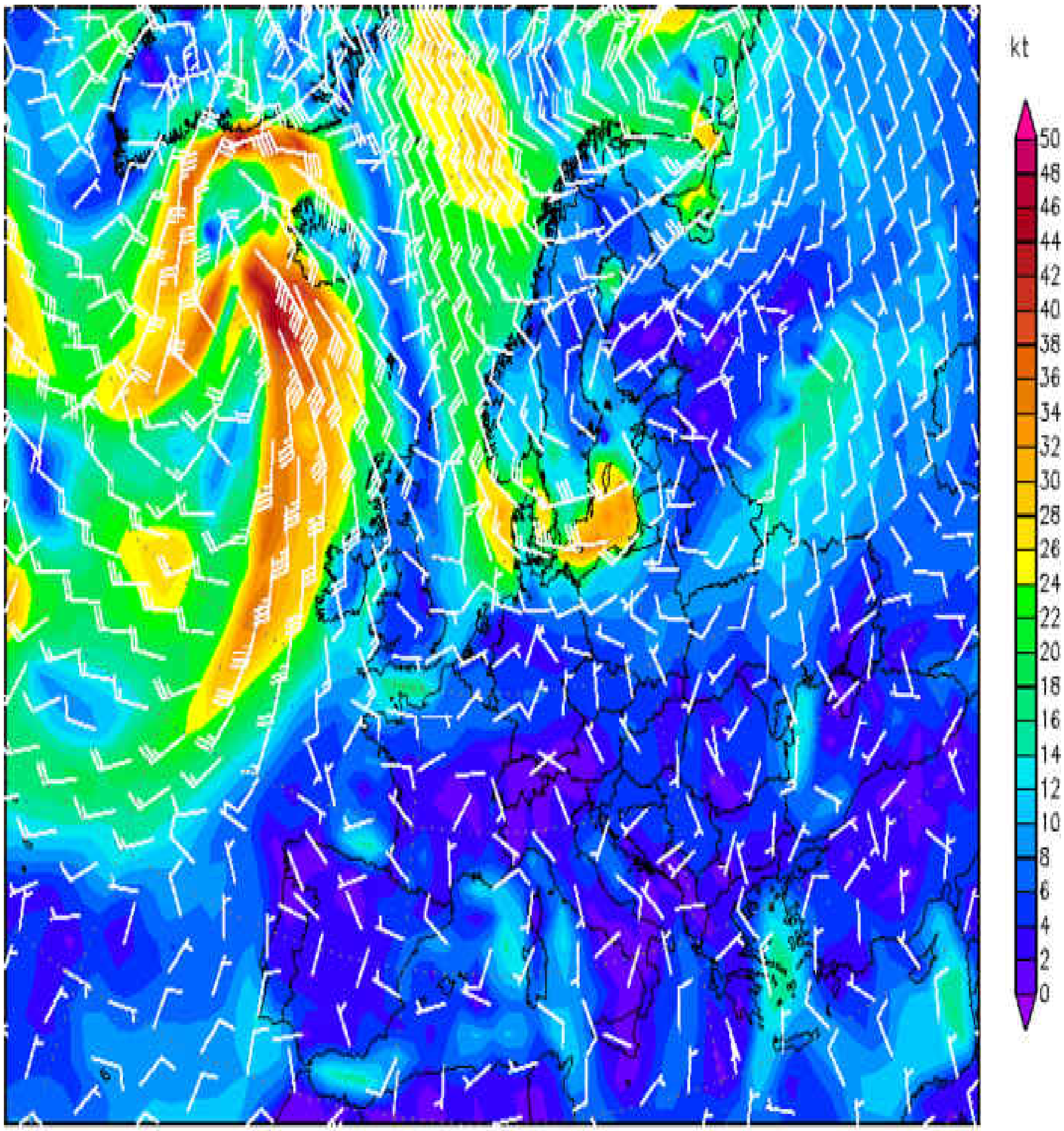

Figure 12.4: Typical wind prediction chart showing contours of wind strength. The barb symbols show direction and local strength (see Figure 12.5).

Wind barb symbols

The features included on these charts tend to differ according to where you get them from. On some charts, whether they are isobar, wind or wave prediction, you might see an additional feature called **wind barb symbols**. These show wind strength and direction at the same time. The barbs are a highly simplified numerical system for wind strength. A full barb represents ten knots, half a barb means five knots, and a large triangular barb means 50 knots (Figure 12.5).

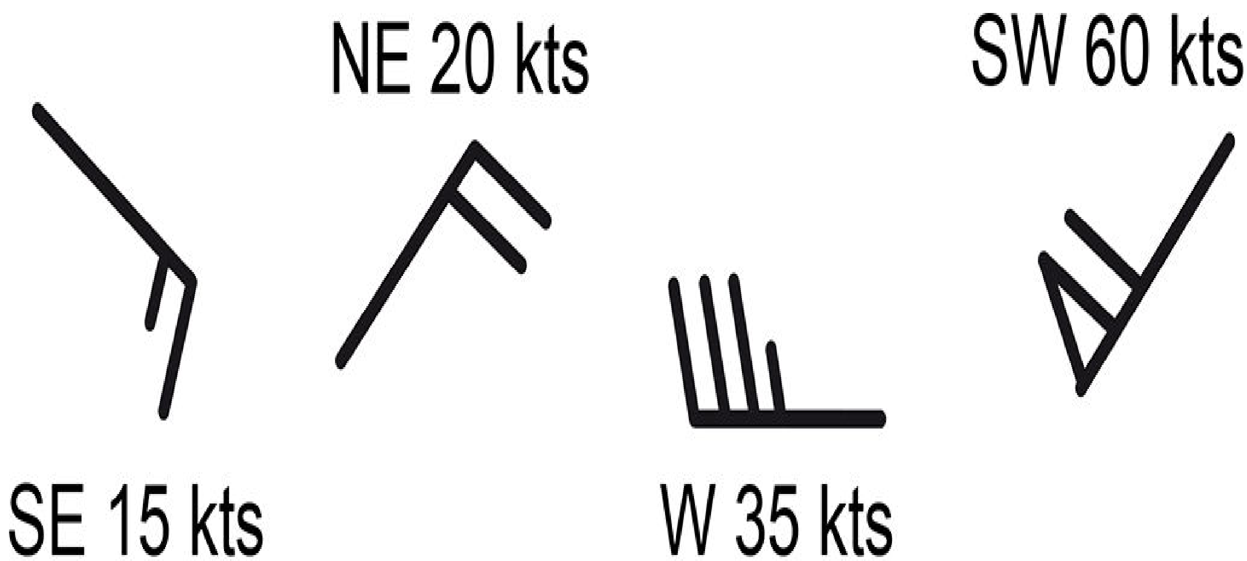

Figure 12.5: Examples of wind barb symbols. Each long barb is ten knots, a short barb is five knots and a triangle is 50 knots.

Wavebuoy reports

To obtain real, measured information on sea conditions a short time ago, we can use the data collected by **wavebuoys**. These sit in the ocean and constantly measure a range of parameters, the most important for us being wave height and period, wind speed and direction. Some kinds of buoys measure wave direction, which is also very useful to us.

Some people get confused and think that wavebuoys can do predictions. It is important to stress that this is not the case. All they can do is make real-time measurements using the different instruments they carry on board. From these measurements you can see the current sea state and its evolution over the previous 36 hours or so. Although the buoys are constantly making measurements, the reports we see are normally made at hourly intervals, and take around an hour (a surprisingly short time) to become available on the internet. They are useful to us if we want to see whether the wave height just off the coast is picking up or dropping off. If the wave height is increasing rapidly at the buoy and you can tell that the swell is headed for the coast, there will be waves on the beach within a short time. Also, a rapid increase in period at the buoy indicates that the first waves in a new swell have already passed that point and might be heading for the coast.

Due to the effect on the waves of shallow water between the buoy and the coast, the exact wave height on the coast will probably differ from that at the buoy. Experience will give you a good idea of how the continental shelf affects the waves propagating in towards your local spot from the nearest buoy. The wave height reduction (which it usually is) differs not only from one spot to another, but also according to the swell quality and direction.

About wind and wave models in general

All wave and wind forecasts, including wave-height and period contour charts, wind charts and those site-specific forecasts we all use every day on surf-prediction websites, come from a mathematical model. This is just a set of mathematical equations, made into a computer program and used to calculate the future evolution of a set of parameters based on their present value. Every prediction in oceanography and meteorology uses a mathematical model.

The wave model calculates what is known as the **directional spectrum**. This is basically the height, period and direction of all the waves passing through a single point on the ocean at a single point in time. Don't worry too much about the details at the moment: the directional spectrum will be covered much more thoroughly in Chapter 13.

The wave model calculates the directional spectrum for each point in an imaginary grid covering either the whole globe or just a section of the ocean. The points on the grid are typically a few kilometres apart. The model then repeats this calculation for a 'target' time – say, a few minutes later. The second calculation is a prediction based on the first calculation. It keeps doing this until it has predicted the directional spectrum for all the thousands of points in the grid, every few minutes for up to, for example, ten days ahead. This is called a model 'run', and the computer does all this very quickly (computers are extremely good at repetitive tasks). A 'run' is usually performed several times a day.

With all that data to process so quickly, powerful computers are needed. This is especially important for short-term forecasts where the computer might not spew out its answer until very close to the actual 'target' time. It would obviously be no good if the computer took so long that the forecast wasn't ready until the target time had passed.

To construct the contour plots of height, period and direction, or to provide those values for site-specific forecasts, single values are extracted statistically from the directional spectrum. This reduces the data down to one value of height, period and direction for each grid point, which is a lot less to deal with. To make the contour charts, the values at each grid point are **interpolated** to produce smooth lines between grid points.

The data for wind prediction charts and isobar charts are derived from a mathematical model of the atmosphere which, just like the wave model, calculates the values of a set of parameters at each point on a world-wide grid. This is also repeated for different 'target' times into the future, using the previous set of values as a starting point. The major difference between the atmospheric model and the wave model is that the atmospheric model 'domain' is three-dimensional, whereas the wave model domain (the surface of the sea) is only two-dimensional. To accommodate this, the atmosphere is divided into a number of layers in the model.

The wave model relies heavily on the atmospheric model, because it uses the wind predictions of the atmospheric model for its inputs. It also uses other variables, such as the water temperature and the position of the ice shelf, together with ‘fixed’ data such as the sea-floor bathymetry and the location of islands and shoals. But the wind inputs are by far the most important. Any slight inaccuracies in the output of the wind model can be amplified in the wave model, resulting in wildly differing wave-model outputs.

The quotation of wave height and period

As mentioned above, in a wave model, a single height and period are extracted statistically from the directional spectrum generated by the model. The way these numbers are derived is intended to reflect as closely as possible what would be observed by eye. The term **significant wave height** (**Hs** or **H1/3**) was originally defined to be compatible with human observation. In the 1940s, Harald Sverdrup and Walter Munk defined significant wave height as the average height of the highest one-third of all waves, and stated that it was ‘about the same as the most common height estimated by an experienced observer’.

While wave height is nearly always quoted as Hs, period may be quoted in several different ways. The **peak period** is the period where most of the energy of the sea is contained. For example, if there are big waves with periods around 12 seconds, but smaller waves in the periods either side of that, it means that most of the wave energy is concentrated around periods of 12 seconds, and so the peak period will be 12 seconds. The **average period**, which you might see on wavebuoy reports, is not quite the same as peak period, but it can be used in more or less the same way. For example, a lot of energy at short periods means both a low-peak period and a low-average period. It is vitally important to know which one you are dealing with, and to only compare like with like.

On the general reliability of predictions

When making predictions of any kind, we must be careful that what we say has at least some chance of coming true. The prediction must have a degree of reliability. Vague statements about the future are more reliable than more specific ones, because they have a wider scope for error. Therefore, if we are trying to predict quite far ahead into the future, the statement we make about the future must not be too specific. Short-term predictions can be more detailed and specific. Here are a few examples:

> 'In two minutes time the sea will probably contain liquid.' *Maximum reliability. This statement is extremely vague and extremely short-term.*
>
> 'There will be some kind of surf this winter.' *Long-term but highly vague, therefore quite reliable.*
>
> 'Tomorrow there will be two- to three-metre waves with an offshore wind.' *Quite specific, but short-term enough to be fairly reliable.*
>
> 'Not next Tuesday but the next, the surf will be one and a half metres and the wind will be offshore.' *Unreliable. This prediction is too far ahead for the details you are giving.*
>
> 'This afternoon, at twenty past three, the wind will turn onshore.' *Unreliable. A very short-term prediction, but still too detailed to be reliable.*

On the whole, wave and weather forecasts are more reliable if the general meteorological situation is stable rather than unstable. If the isobar chart is dominated by a large, slow-moving high-pressure system, then fairly reliable predictions can be made quite a few days ahead.

In contrast, if the chart contains fast-moving and fast-deepening depressions, then things will only be reliable on a short-term basis. In the North Atlantic, the 'stability' of the situation, and therefore the reliability of the forecasts, is linked to a phenomenon known as the **North Atlantic Oscillation** (**NAO**). The **NAO index** is based on the pressure difference between the north and south of the North Atlantic, which oscillates up and down on various different time-scales from a few days to tens of years. If the NAO index is high, namely the pressure is much lower in the north of the North Atlantic than in the south, the North Atlantic is ‘mobile’ and unstable. In this situation, predictions are generally more difficult than the more stable situation of a slow-moving or stationary high pressure in the middle of the North Atlantic, resulting in a low or negative NAO index. There is a lot more about the NAO and surf-related climatic cycles in my book *The Surfer’s Guide to Waves, Coasts and Climates*.

13 Forecasting 2: Inside the Black Box

Introduction

In Chapter 12 I mentioned that the most popular surf-forecasting websites (at the time of writing anyway) give single values of wave height, wave period and direction, plus wind strength and direction, at thousands of specific points along the coastlines of the world, including your local beach. Some of them combine these data with ‘star-rating’ systems that tell you how good the surf is going to be on a scale of, say, one to five. The numbers are seemingly drawn out of a ‘black box’ and presented to us on a plate. As long as this works, most of us don’t want to complicate life by seeing how those numbers were calculated. We are not interested in looking inside the ‘black box’.

However, as I pointed out earlier, if the surf doesn’t turn out the way ‘they’ say it is going to, we wonder why. For example, on your beach, a 6-foot, 15-second north-west swell with a light south wind (or whatever) might usually be clean, lined up and perfect. But today it turns out to be mixed up, choppy and far from perfect, for no apparent reason. In such cases, we need more information than a simple wave height, period and direction.

There are several things we can do, including carefully inspecting the isobar charts, wave prediction charts and wind charts, as described in Chapter 12. If necessary, we can also look at the ‘raw’ outputs of the wave-forecasting model *before* those statistical summaries are calculated. This is the directional spectrum mentioned in Chapter 12 – the raw output from the wave prediction models before the data is reduced down to those statistical summaries.

In this chapter, we will take a look inside that black box. First, I’ll describe how the total sea state is made up from all those waves of different heights, periods and directions mixed together. Then I’ll introduce some plots of the directional spectrum and give examples of how to interpret them.

Real waves aren't all the same

Waves of a single height, period and direction do exist. They are called **monochromatic** – of a single period – and **unidirectional** – of a single direction – and can only be found in laboratories and textbooks. The real sea contains a large number of heights, periods and directions. Even the most perfectly groomed swell arriving at the coast from a storm thousands of miles away is still never totally regular.

In Figure 13.1 you can see two different wave trains. The diagrams are intended to simulate the ups and downs of a floating object on the water surface, plotted as a function of time. The top one is a single sine wave generated in a laboratory or on a computer. It has only one height (2 m) and one period (7 seconds), and is totally regular. The bottom one is a natural wave, similar to what you might see on the surface of the real ocean. It has no easily definable height or period. The diagrams do not show direction, but try to imagine that the top wave is always coming from the same direction, whereas the bottom one is not.

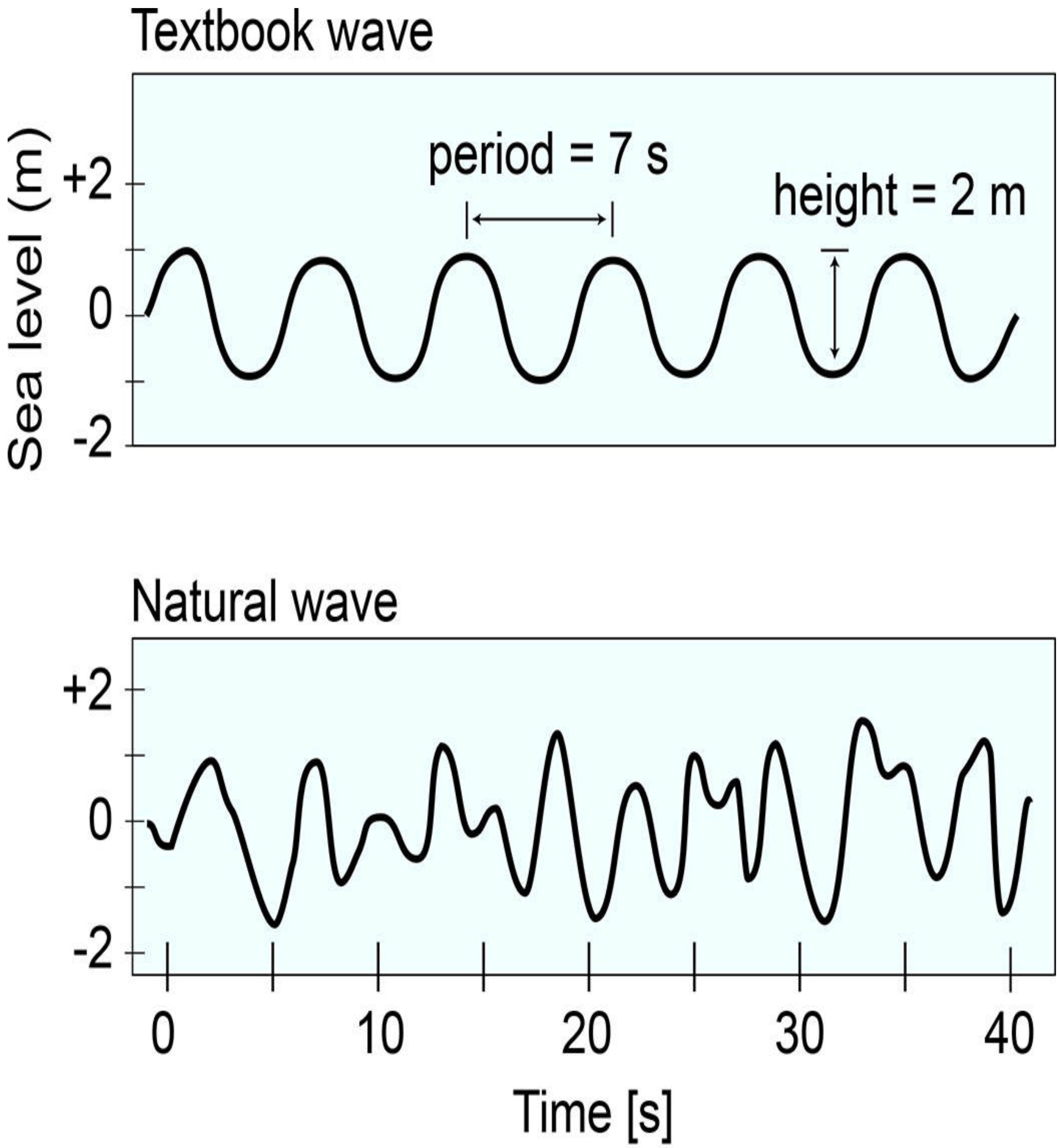

Figure 13.1: A 'textbook' wave (top), and a more realistic representation of the sea surface (bottom).

It follows that we could think of the sea state at any given moment as consisting of not just one ‘wave component’, but several, each with its own particular height, period and direction. These components would have propagated to the point of observation from different regions of the ocean. This is not exactly what happens in real life, but it is a nice simple analogy to help us understand something that is pretty complex.

Decomposing a complex wave

How is a single value of height, period and direction extracted from a wave like the one at the bottom of Figure 13.1? The easiest way might be to decompose the ‘raw’ record of sea-surface ups and downs into all the original components encrusted inside it. Each of these components would be a regular **sine wave** like the one at the top of Figure 13.1, each with its own height, period and direction. Decomposing the wave field into its component sine waves is actually a simple and painless exercise for any computer. Once all these individual waves have been extracted, you could then represent them in a large number of diagrams, such as the one at the top of Figure 13.1.

However, this would be rather clumsy and very difficult to interpret. A much better way is to use a plot of wave energy versus period – a **spectrum**. A spectrum can be thought of as a kind of statistical distribution of all the component waves that go to make up the wave field. By definition, the **peak period** is where the energy is highest. Then, at longer and shorter periods, the energy gradually diminishes. The amount of wave energy at a particular period is governed by the height of the waves around that period. For example, a peak period of 15 seconds indicates more wave energy at that period than at any other, which, in turn means bigger waves with periods of around 15 seconds than at any other period (Figure 13.2).

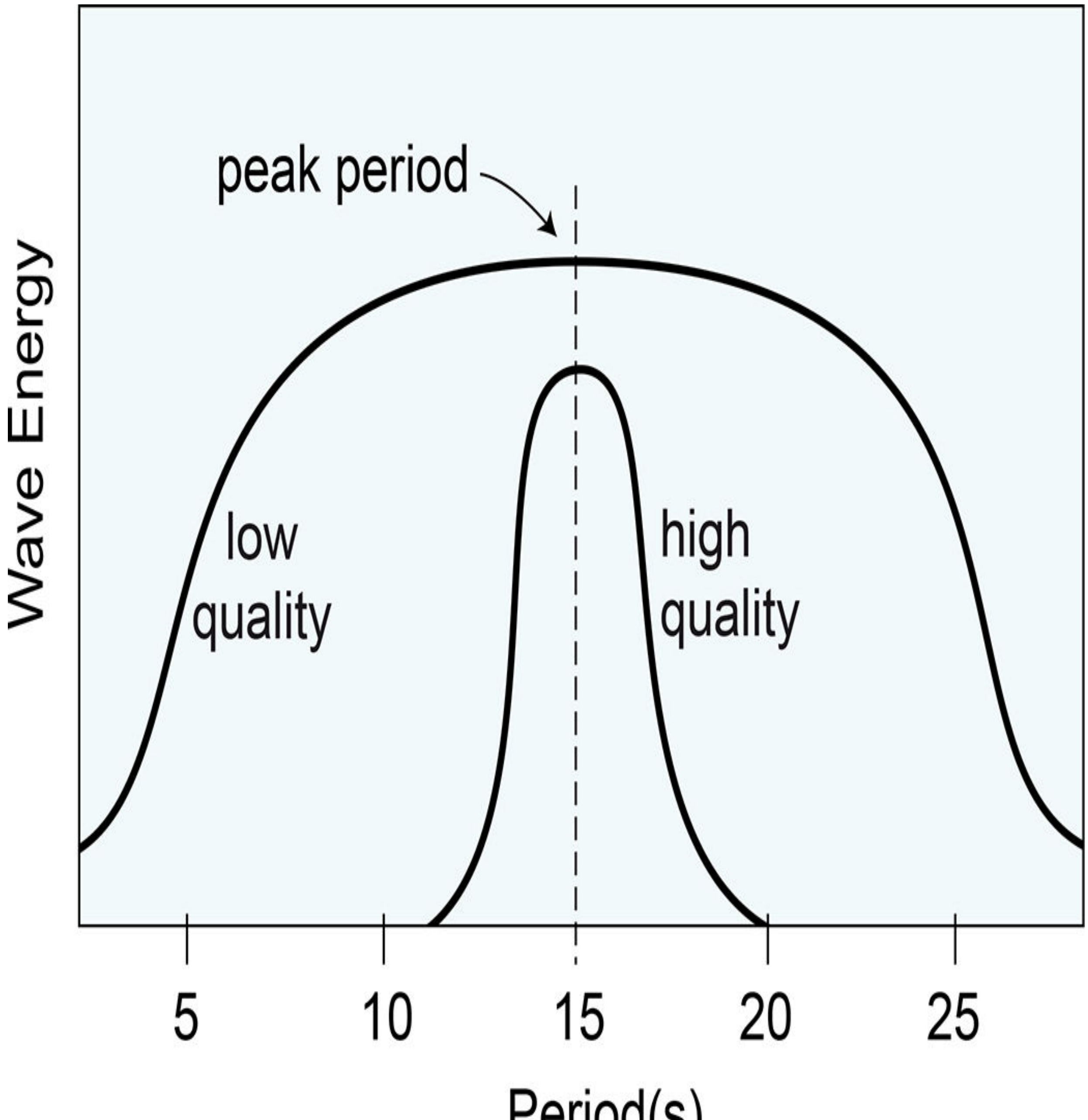

Figure 13.2: A wave spectrum. The broad and narrow curves represent low- and high-quality swells respectively.

Bandwidth

Going back to those statistical summaries, if a single period is quoted, the one most often quoted is the peak period. Looking at Figure 13.2, it is apparent that the peak period only tells us at what period there is most energy. It doesn't tell us anything about the distribution of energy at longer or shorter periods – the **bandwidth**. For example, a wider, flatter curve (broad bandwidth) means that the energy is more spread out over the whole range of periods. Here, the sea will appear messy and confused, because there are just as many big waves at short and long periods as there are at mid periods. In contrast, a narrow, pointed curve (narrow bandwidth) means all the energy is concentrated around the peak period. Therefore, all the big waves are concentrated around a narrow range of periods. This suggests a purer, cleaner swell.

What is crucial about the two curves in Figure 13.2 is that, although one swell is very different from the other, they both have the same peak period (15 seconds). If you only have the peak period, it is impossible to know how the energy is spread around that peak. Two different swells on different days might have the same peak period but they might have completely different bandwidths. As a result, one swell might be messy and confused whereas the other might be much cleaner and lined up.

Directional spread

Just as real waves never contain only one period, neither do they contain only one direction. Like monochromatic waves, unidirectional waves exist only in laboratories and textbooks. In the real ocean, waves are always coming from many different directions at the same time. That's pretty obvious if you have ever been in a boat, and it is usually quite noticeable even when you are sitting on your board during any normal surf session.

The variation in direction of a wave field is analogous to the idea of bandwidth. Even though the waves never all come from exactly the same direction, sometimes they come from almost the same direction, and other times they come from a very wide spread of different directions. If all the wave components are clustered around a small range of directions, then the wave field is said to have a narrow **directional spread**. But if the wave components vary a lot in their direction of approach, then there is a wide directional spread.

A narrow directional spread means you will get a clean swell with straight, ruler-edged lines, because most of the waves are coming from the same direction. A wide directional spread means you will get a peaky, mixed-up swell, because the waves are coming from many different directions and interfering with each other.

In wavebuoy reports and wave-prediction websites, a common way of summarizing the direction statistically is to use the direction from which there is most energy, called the **peak direction**. If you have only the peak direction, then it is impossible to know how much directional spread there is in the wave field. Two different swells on different days might have the same peak direction, but they might have completely different directional spreads. As a result, one swell would be nicely lined-up, the other would be peaky and irregular.

Usually, swells that arrive from distant storms tend to have narrow directional spreads, and swells generated more locally tend to have wide directional spreads. At a particular observation point, swell can only be received from within the ‘field of view’ covered by the storm from the point of view of the observer. A storm thousands of miles from the observer has a relatively narrow ‘field of view’, and hence the swell will arrive at the observation point from only a narrow range of directions. In contrast, a storm that is practically on top of the observation point has a much wider ‘field of view’, allowing swell to arrive from many different directions at the same time (Figure 13.3).

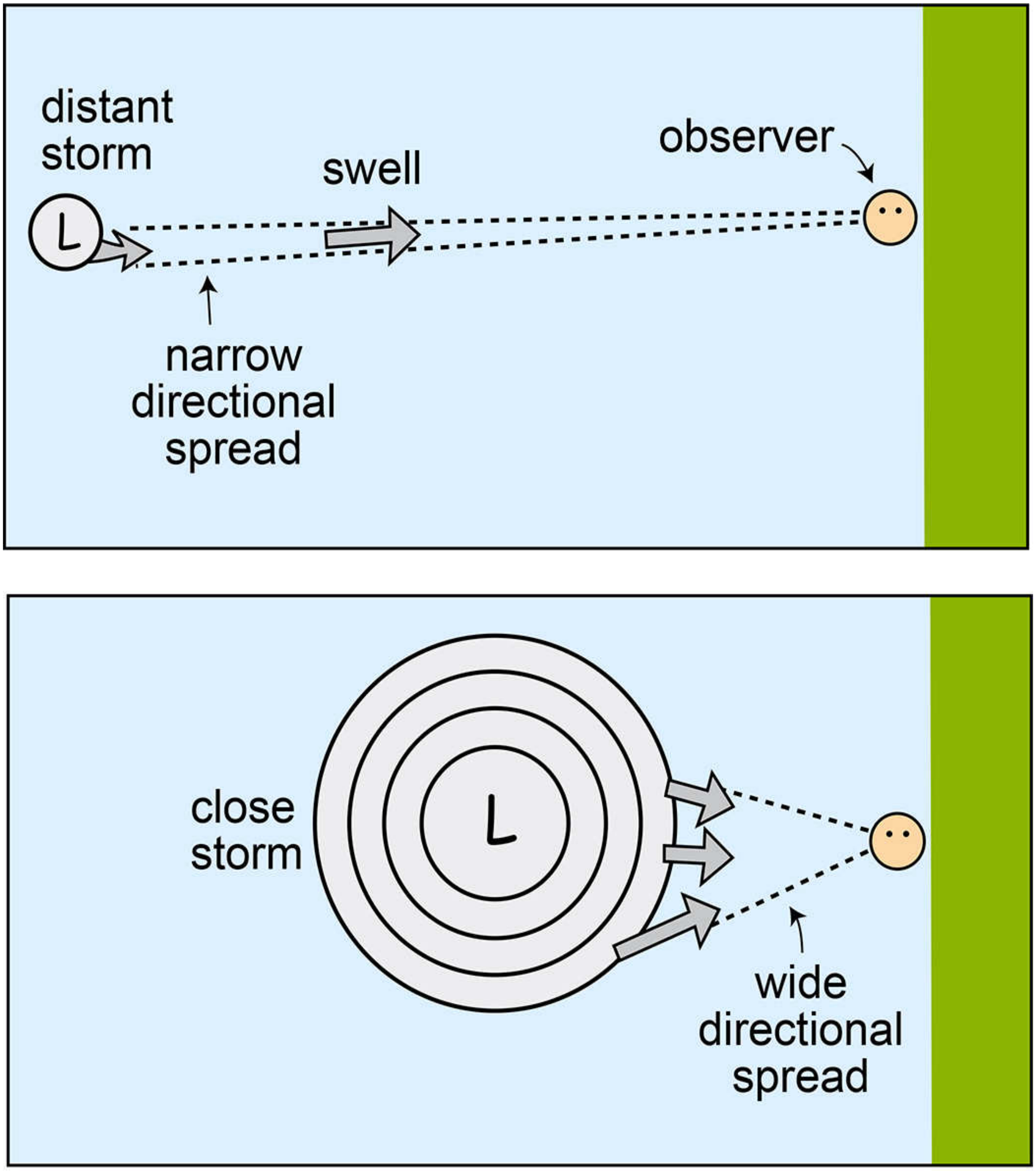

Figure 13.3: Swells coming from distant storms have narrower directional spreads than swells generated close-by.

The directional spectrum

To obtain all that missing information contained in the wave field, either in real time or as a prediction, the wavebuoys or the wave prediction models must have the necessary capability. In fact, many wavebuoys do measure the direction of wave approach at the same time as the ups and downs of the sea surface. As far as the wave prediction models are concerned, most of them churn out the full directional spectrum at every one of their thousands of prediction points around the globe. The problem is that we, the public, only see that information at a select number of points. At those points, the full directional spectrum data are available to view or download from the internet, usually in the form of a three-dimensional plot.

The **directional spectrum plot** is more or less what it says – a spectrum like the one in Figure 13.2, but with wave energy plotted as a function of period *and* direction. It shows us how the energy is distributed over all periods and directions, and tells us not only the peak period and direction, but also the bandwidth and directional spread. Because there are three axes, the plot must be three-dimensional. If you have seen contour plots of mountain ranges, or of sea-floor bathymetry, then this is really not much different. In fact, the contour plots of wave height, wave period or windspeed shown in Chapter 12 are not much different either. The third dimension is simply represented by contours of different colours.

One thing that people find difficult to get their head around is the way the axes are laid out on these plots. But it is quite simple really. Figure 13.4, for example, shows an 'empty' directional spectrum plot. The axis representing period is the radial distance from the periphery of the circle towards the centre, and the axis representing direction is the periphery of the circle itself. Colour contours of wave energy are plotted on the graph, according to the period and direction of that energy. If you are confused, think of it as energy contours plotted on a flat, rectangular graph whose dimensions are wave period and wave direction, but, to make it easier for us to visualize, the flat graph is stretched around into a circle.

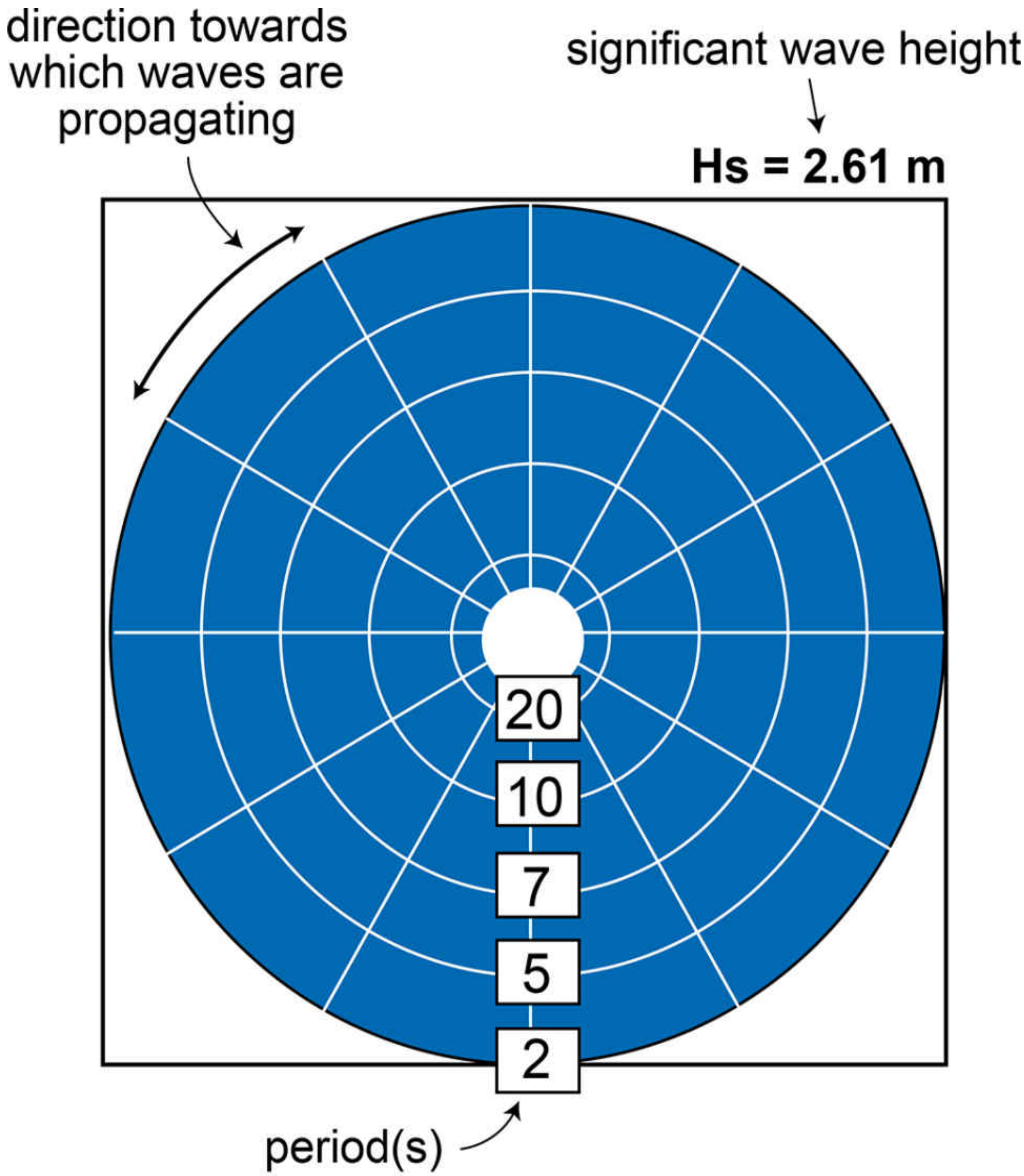

Figure 13.4: Axes of a typical directional spectrum plot. Note that the period is on a logarithmic scale.

The conventions are usually as follows. The direction around the circle is the direction towards which the waves are propagating, and the period increases from the periphery of the circle towards the centre. For example, any energy contours near the centre of the circle mean that this energy was produced by long-period waves, and any energy contours on the right-hand side of the circle mean that the energy was produced by waves travelling from west to east. The significant wave height, which I'll talk about in a minute, is usually given on the diagram as well. Lastly, because scientists like to talk in frequency instead of period (frequency is simply the inverse of period), the period scale ends up logarithmic on these plots, which bunches up the contours towards the centre. This is actually quite inconvenient for us, and could easily be changed.

If you are still confused, look at the following examples and you'll get the hang of it. A 'blob' of energy on the right-hand side of the circle, quite near the centre, means a long-period westerly swell. If the blob is not very spread out (Figure 13.5), it means that both the directional spread and the bandwidth are quite small, and the swell will be of high quality. In contrast, if the blob of energy is really spread out (Figure 13.6), it means a large bandwidth and a large directional spread, resulting in a much poorer quality swell.

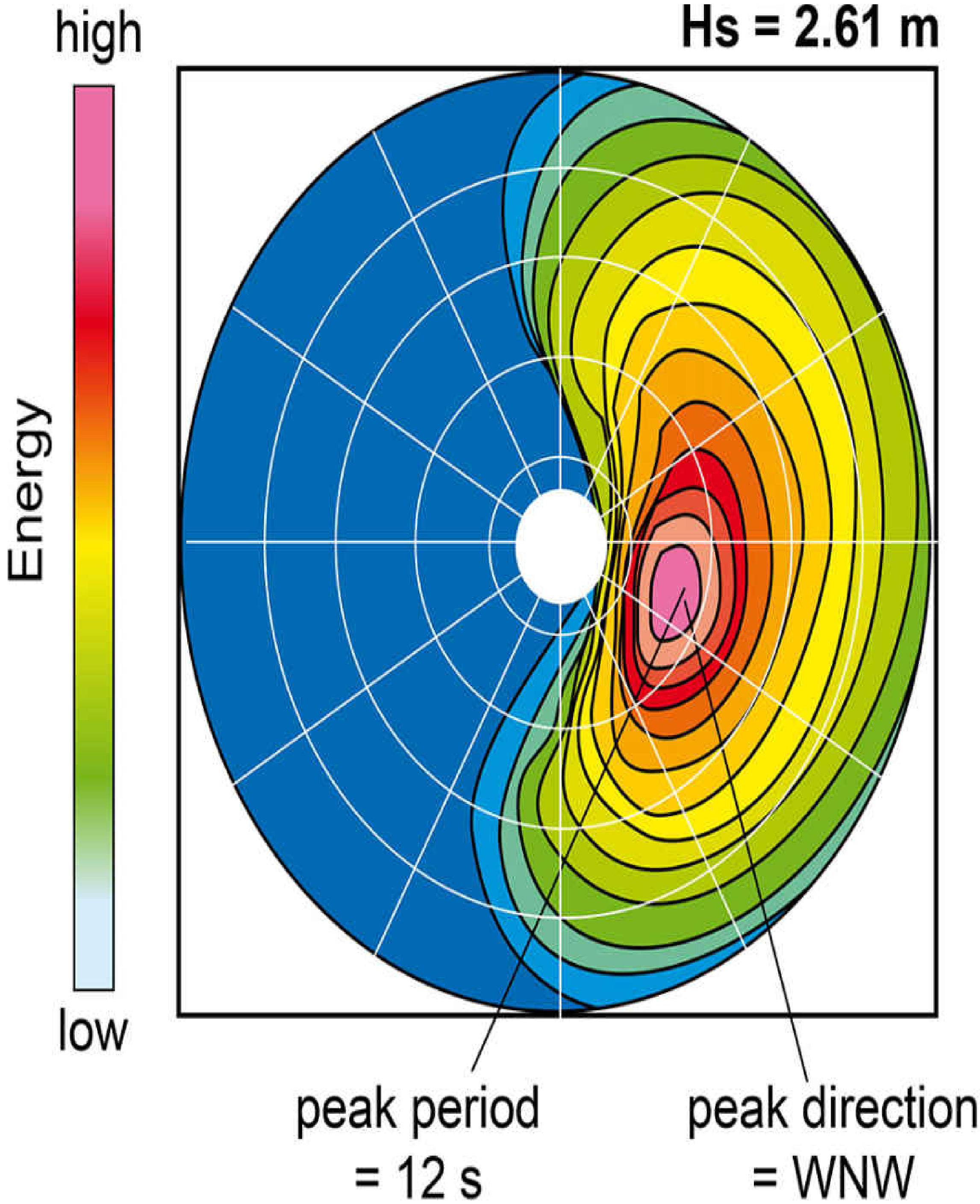

Figure 13.5: Hypothetical directional spectrum plot of a high-quality swell.

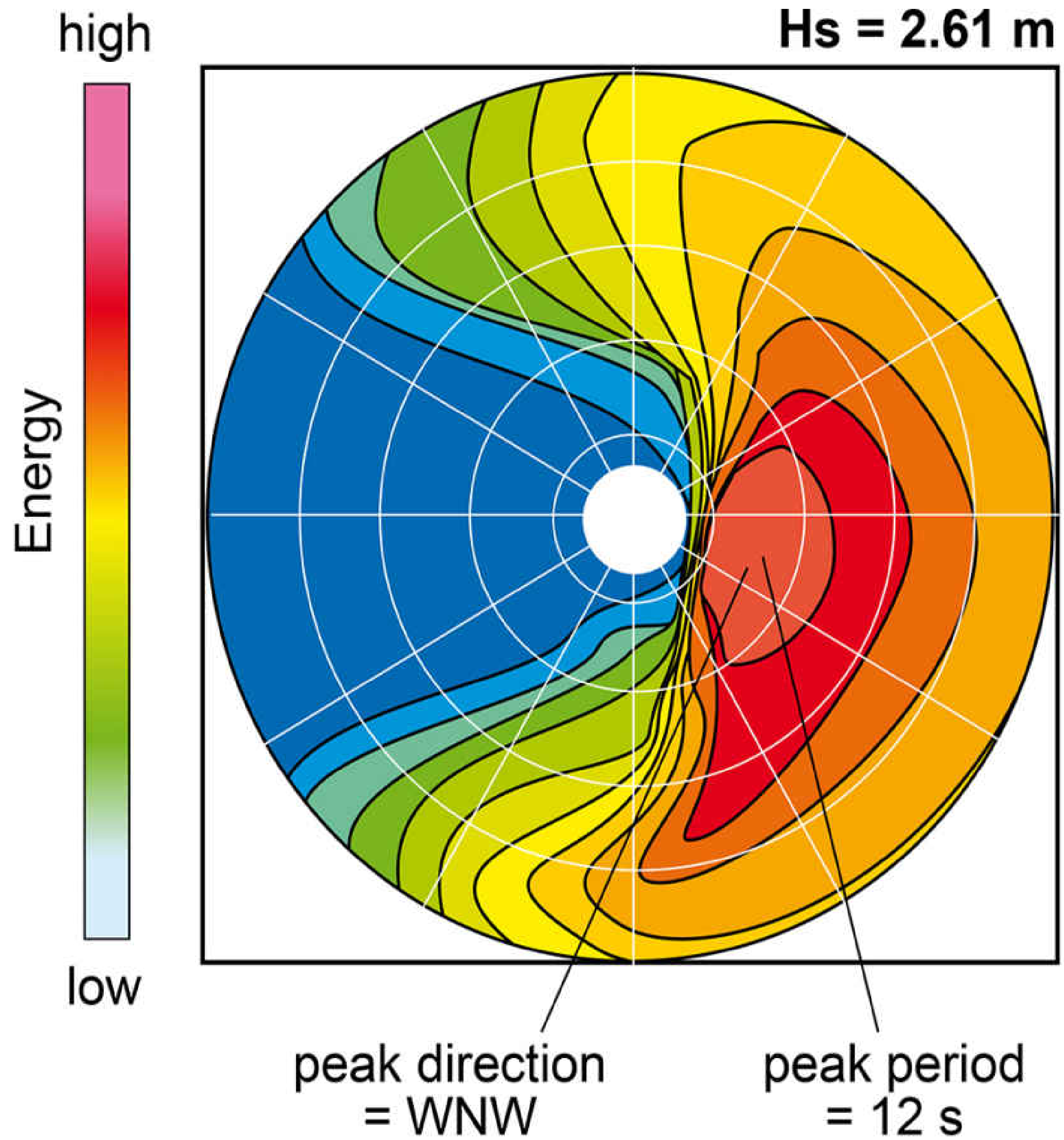

Figure 13.6: Directional spectrum plot of a low-quality swell. The significant wave height, peak period and direction are the same as in Figure 13.5.

What is really important is that the significant wave height, peak period and peak direction (12 seconds and WNW respectively) are exactly the same for the swells shown in Figures 13.5 and 13.6. So, if you were given only these three numbers and no information about the bandwidth and directional spread, you wouldn't be able to tell the difference between them.

Being able to 'look inside' the directional spectrum gives us a wealth of information about the swell. There are many more examples of the usefulness of the directional spectrum apart from just being able to distinguish between good- and bad-quality swells. For instance, single values of height, period and direction would not give you any information about two swells coming in at the same time. Perhaps if you were to go back and carefully examine the isobar charts and identify two storms in the ocean you might be able to crudely track their swells. Or, perhaps, if you looked at the wave-height contour charts over the last few days you might be able to do the same thing. But with the directional spectrum plot, the presence of two or more separate swells shows up unmistakably. Figure 13.7 shows NW and WSW swells arriving at the same time – quite common on west-facing coasts in the Northern hemisphere.

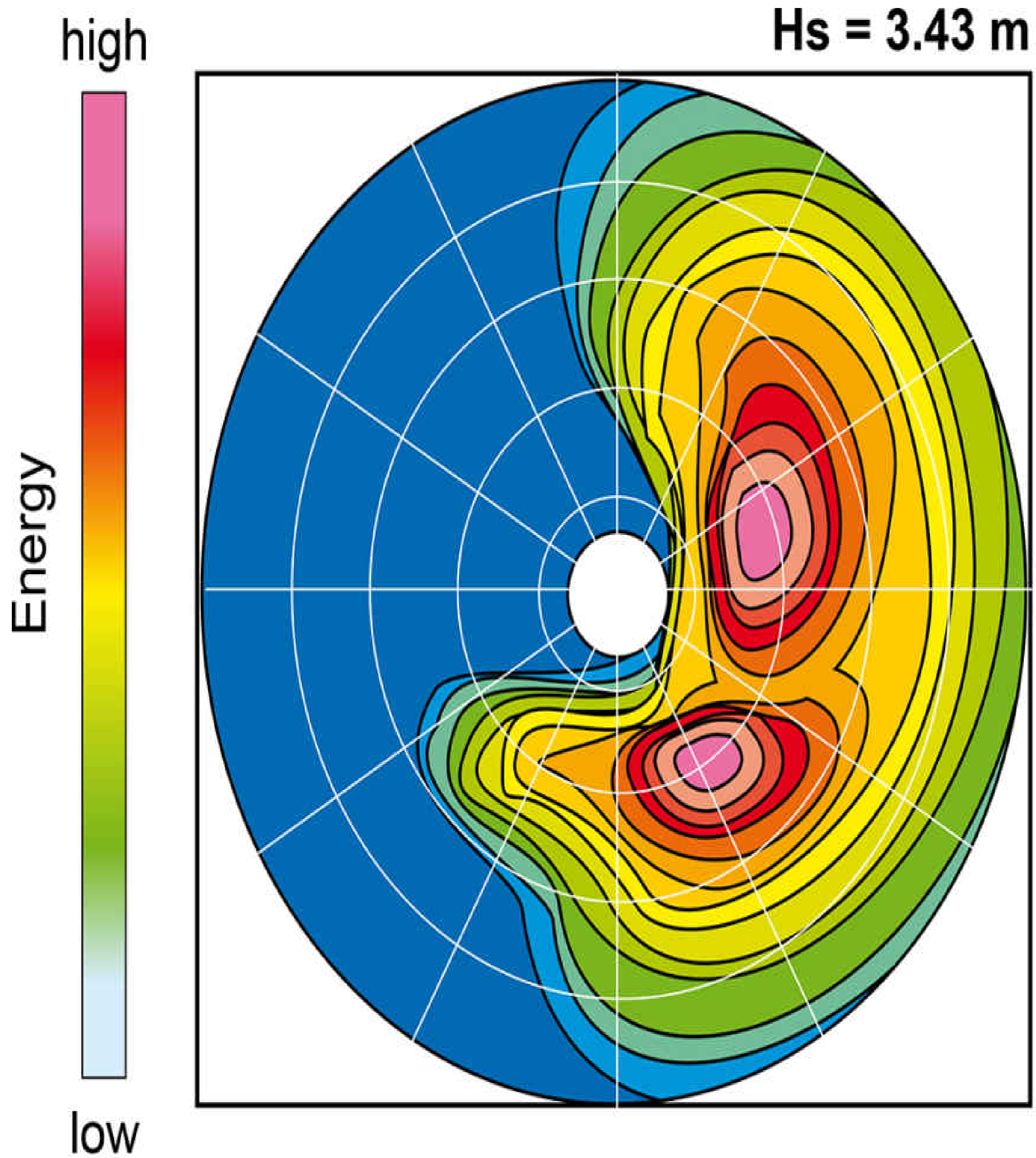

Figure 13.7: The directional spectrum plot can clearly show two or more separate swells at the same time.

Examining the directional spectrum not only enables us to distinguish clearly between long-travelled swell and locally-generated windsea, it can also tell us a lot about the interaction between the two. Figure 13.8 is a plot taken from a point a short distance (say, 100 km) out to sea from a west-facing coast. It shows how a strong offshore wind can remove the short-period waves from a swell.

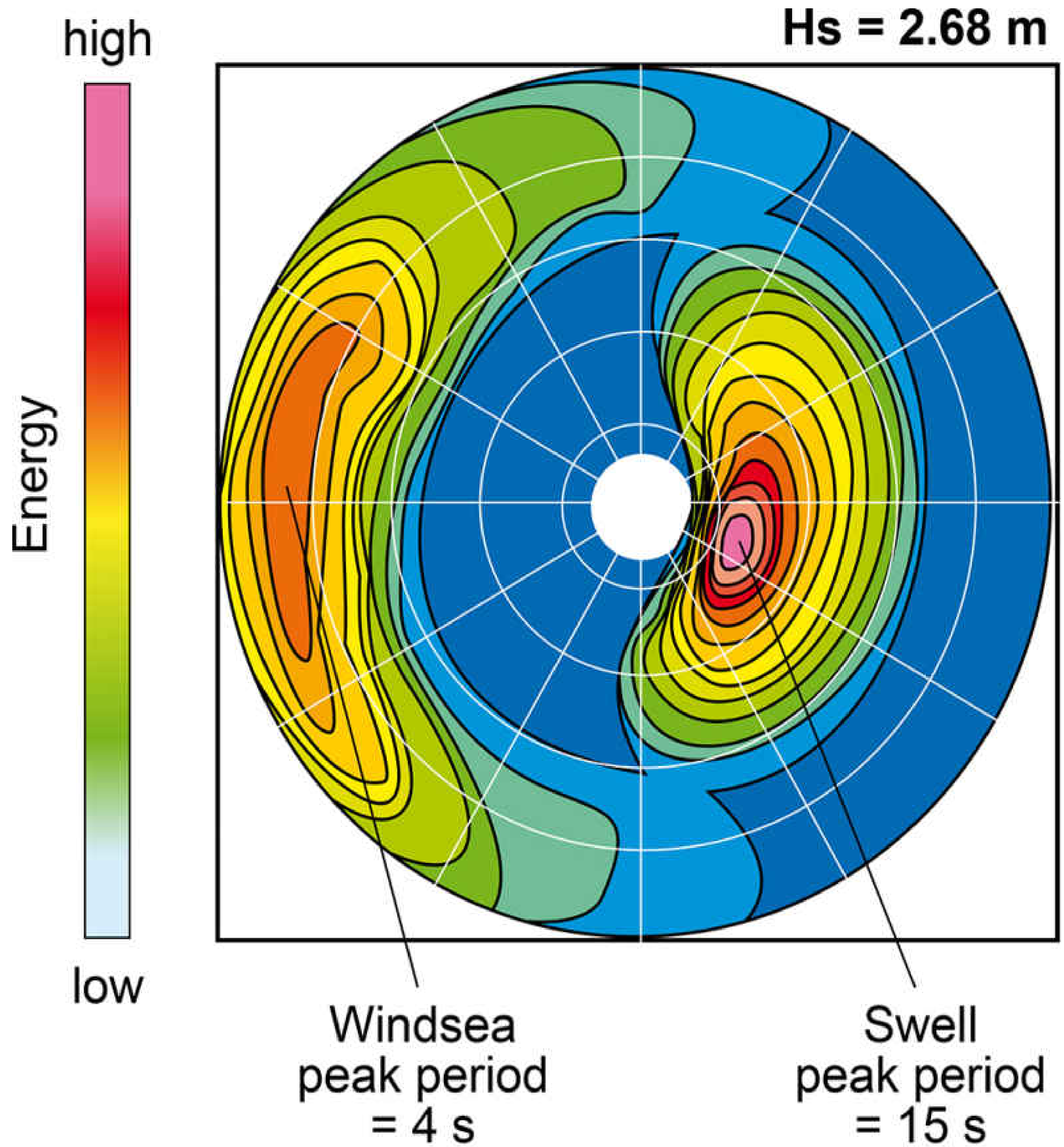

Figure 13.8: An offshore wind cleaning up a swell by removing the short-period waves. The main swell is on the right of the plot and the windsea due to the offshore wind is on the left.

The contours on the right-hand side of the plot represent a clean WNW swell, similar to that in Figure 13.5, but with all the short-period waves (near the periphery of the plot) removed. At the same time, there is a very poor-quality easterly windsea being generated by the offshore wind itself, showing up as the broad band of contours on the left-hand side of the plot. This clearly demonstrates why an offshore wind tends to clean up a swell. I will talk more about this in Chapter 14.

But where does the wave height come in all of this? How can we tell from the directional spectrum how big the waves are going to be? You can't just tell the height by looking at the directional spectrum. So we must still rely on a statistical summary: the well-known **significant wave height**. In theory, the significant wave height should correspond to the height an 'experienced observer' might guess the waves to be. Because even 'experienced observers' tend to focus their attention on the biggest waves, the statistical parameter found to be closest to the observed wave height was originally calculated from the average of the highest third of the waves observed over a certain length of time. Nowadays, the significant wave height is based on a kind of average of all the energy in the directional spectrum, multiplied by a 'boost' factor to compensate for the observer's bias towards the highest waves.

Note that the energy contours are on a 'sliding scale'. The contours give only the *relative* distribution of energy over period and direction, not absolute values of energy. A bright blob shows only where the maximum energy is compared with all the other energy on the plot. If we want to get some idea of how big the waves are actually likely to be, we must use the significant wave height to *calibrate* the plot. The reason that the contours are on a 'sliding scale' is because it allows the full range of contours to be used no matter what the absolute values of energy or wave height are.

14 Forecasting 3: Surfing in the Storm

Introduction

The first seven chapters of this book described a situation in which a low pressure is spawned somewhere in the middle of a large ocean. The waves generated here propagate thousands of kilometres before they reach some coast, where we ride them. We discovered how, after propagating some distance away from the storm centre, a whole mixture of different waves can be converted into clean, ruler-edged lines, perfect for surfing.

But what about those places where there's just not enough room for a large storm to develop and propagate its waves thousands of kilometres? Many surfers around the world do not have the luxury of long-travelled swells. Instead, they must wait for a weather system to pass right over the top of them, and for the wind to blow onshore long enough to generate some kind of rideable surf. Then, for the surf to clean up, they must wait for the wind to turn offshore. Here, people surf right in the middle of the storm itself. The waves have had no chance to propagate away as free-travelling swell.

The diameter of a typical oceanic low-pressure system is around several hundred kilometres. Add to this a few more hundred kilometres for the swell to propagate, and you have quite a large area required for clean swells to develop. In some places, such as the English Channel, the Mediterranean, the Great Lakes and the Gulf of Mexico, there is just not enough room for this to happen. Of course, swells can occasionally enter some of these areas from the open ocean, but most of the time, the surf is generated by storms that end up right on top of the surf spot itself. If you want to surf here, first you have to wait for a weather system to pass right over, and for the wind to blow onshore long enough to generate some kind of surf. It will be a nasty, chopped-up windsea. Then the wind might switch offshore and you'll get clean, rideable surf for a few hours. The trick is to be on it. Hit the wrong spot at the right time, or *vice-versa*, and you've blown it.

A typical case might be on the south coast of England, which is located in the direct path of North Atlantic winter storms. As a low-pressure system approaches from the west, the wind is onshore from the south or south-west, blowing over a distance limited by the width of the English Channel. As the storm passes over, this wind veers round to the north-west. Tell-tale signs on a weather chart include a cold front connected to a centre of low pressure somewhere to the north. In front of the front is warm air blowing from the south-west, and behind the front is cold air coming from the north-west (Figure 14.1).

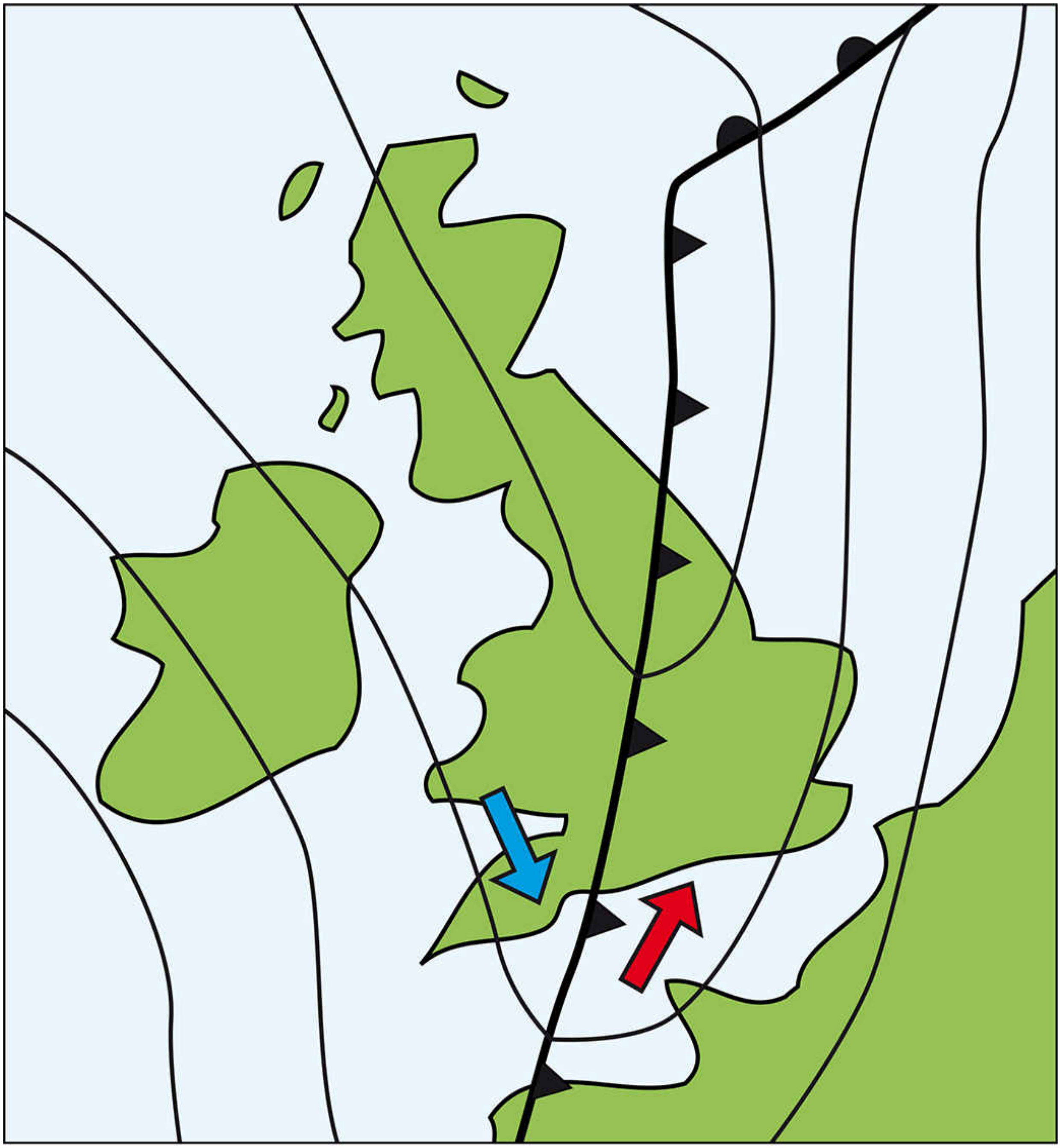

Figure 14.1: Typical weather chart for the south coast of England: As the cold front passes from west to east, the wind veers from south-west to north-west.

As the front itself passes over, the wind normally goes calm, and there is heavy rain. So it's pretty easy to recognize the situation as it unfolds in real-time. The same sort of scenario plays out in hundreds of other places around the world, although the orientation of the cold front and wind directions might be different.

Surfing in the storm requires local knowledge and short-term prediction techniques. Often, the surf in these places is more difficult to predict than the usual long-travelled swells that take several days to arrive. Instead, waves come and go on an hourly basis, and conditions change very rapidly.

In the rest of this chapter we'll look at a few of the peculiarities of surfing in the storm. I'll examine typical scenarios of waves generated by local winds, the maximum height the waves can reach and what happens when the wind switches direction. In order to do this, we'll use a 'manual' method of wave prediction. This method, although mostly used for academic purposes nowadays, is quite versatile in situations like this – and you don't need a supercomputer. It is also a good excuse to go back in time and have a look at one of the very first wave-forecasting models invented.

The SMB model

Logically, the longer the wind blows from the same direction (the *duration*), the bigger the waves will be. Also, the waves will have a greater chance to grow if the wind blows over a longer distance (the *fetch*). These two factors combined with the wind strength itself are the three essential parameters that determine the height to which waves can grow in the storm. For surfing in the storm, we will use a simple wave-forecasting model based on these three parameters.

The model was developed by Harald Sverdrup and Walter Munk around the late 1940s, with some later contributions from Charles Bretschneider. The model is a set of equations into which you put values of windspeed, fetch and duration, and out of which come single values of wave height and period. In other words, the height and period of the waves are dependent upon the fetch length (longer fetches produce bigger waves), the duration (the longer the wind blows, the bigger the waves) and the windspeed itself (the stronger the wind, the bigger the waves). It is called **SMB theory**, after its inventors. It only considers the height and period of the waves in the storm itself, but not after they have started to propagate away as free-travelling swell.

Apart from occasional use by coastal engineers who want a rough estimate of maximum expected wave heights when designing a pier or other coastal structure, nowadays SMB theory is mainly used only for academic purposes. However, its versatility and simplicity make it a great tool to use for the simple example we will look at here.

The SMB model was based on mathematical relationships between the wind and the waves, but not necessarily fully based on the real physics behind wave generation. It was developed by collecting measurements of windspeed, fetch and duration in a storm, together with the corresponding wave height, and repeating these measurements a vast number of times for as many different situations as possible. After the data were collected, graphs were drawn up and analyzed to find out how wave height was related to windspeed, fetch and duration. What resulted was a set of equations. You don't have to worry about the equations themselves – it is more important just to know that there are three of them, each representing one of three different cases: **fetch-limited**, **duration-limited**, and **fully arisen sea** (Figure 14.2).

Fetch-limited:

$$\boxed{\text{height}} = \boxed{0.0045 \times \text{wind} \times (\text{fetch})^{1/2}}$$

Duration-limited:

$$\boxed{\text{height}} = \boxed{0.00355 \times (\text{wind})^{5/4} \times (\text{duration})^{3/4}}$$

Fully arisen sea:

$$\boxed{\text{height}} = \boxed{0.00192 \times (\text{windspeed})^{2}}$$

Figure 14.2: SMB equations. Wave height is in metres, wind speed in km/h, fetch length in km and duration in hours.

… In the fetch-limited case, the waves will only get bigger if the fetch gets longer or the wind gets stronger, but any increase in duration won't make any difference. This normally happens in short-fetch situations such as the English Channel example shown in Figure 14.1.
… In the duration-limited case, the waves will only get bigger if the windspeed increases or the wind blows for longer, but any increase in fetch won't make any difference. This normally happens at the beginning of a storm when the wind has only been blowing for a very short time.
… In the fully arisen sea case, the waves have reached their limit for that particular windspeed. It doesn't matter how much you increase the fetch, or for how long the wind keeps blowing, the waves won't get any bigger unless the windspeed itself increases. This normally happens in situations where the wind has been blowing for very long periods of time over very long fetches.

Before using the equations, you won't know whether the sea state is fetch-limited, duration-limited or fully arisen, so how do you know which equation to use? Because they are *limiting* states, you must first calculate the wave height and period using all three sets of equations, and then take the *smallest* values. Perhaps the easiest way to understand all this is to look at a few examples:

Case 1: Duration limited
A wind of 40 km/h has been blowing over a fetch of 300 km for three hours. Putting those values into each of the equations and calculating the wave height for each case, it turns out that the smallest value for wave height is 0.8 m, which corresponds to duration-limited conditions. If the wind now blows for another three hours (a total of six hours), and put that new value back into the equations, we find that the wave height has increased to 1.4 m.

Case 2: Fetch-limited
A wind of 40 km/h has been blowing over a fetch of 100 km for 48 hours. Putting these values into the equations, we find the smallest wave-height value is 1.8 m, corresponding to fetch-limited conditions. Another day, a 40 km/h wind blows from a different direction, this time over a distance of 200 km, for 48 hours. If we put those values back into the equations, we find that the wave height is now 2.5 m.

Case 3: Fully arisen sea
A 40 km/h wind has been blowing over a fetch of about 4,000 km for at least ten days (= 240 hours). If we put these values into the equations, we find that the smallest wave-height value is 3.2 m, corresponding to a fully arisen sea. This is the maximum wave height possible for a windspeed of 40 km/h. If the wind were to increase to, say, 50 km/h, the wave height would increase to 4.8 m. Increasing the fetch or duration won't make any difference.

The concept of a fully arisen sea implies that, for each wind speed, there is a maximum height to which the waves can grow. In a fully arisen sea, the wave height goes up much faster than the corresponding windspeed; in fact, every doubling of the windspeed means a quadrupling of the wave height.

Most real examples of surfing in the storm usually turn out to be fetch-limited. Duration-limited and fully arisen seas are normally found out in the open ocean, where you have swells arriving from distant storms at the same time as locally generated windsea. In these cases, predicting the waves with any great accuracy is too complicated for a simple model like the SMB.

The effect of an offshore wind

The direct passage of a storm over the place where you are surfing often means that the wind will switch from onshore to offshore, sometimes quite suddenly if a cold front goes over. In western Ireland, for example, which is in the direct path of North Atlantic low-pressure systems, the wind can go from 50 km/h onshore to 50 km/h offshore in less than half an hour.

Before the wind changes, the sea contains a spectrum of different sizes, directions and periods, mixed up together (a typical windsea has a broad bandwidth and directional spread – see Chapter 13). The shorter-period waves in the spectrum are less useful to us for surfing than the longer-period ones. Most of the time, the short-period waves chop up the sea surface, and make things uncomfortable for surfing. If they could be removed, the sea surface would become smoother, the longer-period waves would become more discernible, and the surf would become cleaner.

Fortunately, that is exactly what an offshore wind does – it removes the short-period waves. The shorter waves tend to be steeper, and so ‘stick up’ more out of the ocean which, combined with the fact that they are slower and less powerful, makes it easier for the offshore wind to remove them. The offshore wind has less effect on the longer waves because they are less steep, faster and don’t ‘stick-up’ out of the ocean so much. In Figure 13.8 there is an excellent example of a directional-spectrum plot showing how an offshore wind has removed the short-period waves, leaving only a high-quality swell.

Further Reading

If you would like to delve a little further into any of the subjects I have covered, then have a look at some of the books listed below. There are thousands of books on oceanography, meteorology and waves – these are just the ones I have had the best experience of. The main problem with books about waves is that they tend to get very mathematical very quickly. So I have divided the list of books into four categories, helping you to steer away from unsuitable ones if you are not particularly mathematically minded.

Surfing books containing some basic science

bbott, R., 1972. *The Science of Surfing*, John Jones Cardiff, 91 pp. Interesting early look at surfing from a scientific point of view, mainly UK based.

ordham, M., 2008. *The Book of Surfing*, Bantam Press, 288 pp. Superbly illustrated and stylishly written guide to almost everything about surfing, including travel, history, style and culture, with several pages on the basic science of waves.

rigg, R., 1998. *Big Surf, Deep Dives and the Islands: My Life in the Ocean*, Editions Limited, 179 pp. Nicely illustrated autobiography of Ricky Grigg, big-wave pioneer and oceanographer at the University of Hawaii, including some useful basic oceanography principles.

ampion, D., 1997. *The Book of Waves*, 3rd edn, Roberts Rinehart, 180 pp. Classic book describing waves with beautiful colour illustrations and photos, and a few pages on the formation and propagation of waves.

lorris, V. and Nelson, J., 1977. *The Weather Surfer*, Grossmont Press, 103 pp. Another early guide to meteorology and oceanography from a surfer's point of view, US based.

utherland, B., 2008. *The Stormrider Surf Guide Europe*, 4th edn, Low Pressure, 399 pp. Comprehensive guide to the surfing spots of Europe containing excellent maps and photos, and a basic section

on wave prediction. Many other ‘Stormrider Guides’ are available, most of which include basic meteorology sections.

Textbooks: not at all mathematical

hrens, C. D., 2008. *Meteorology Today: An Introduction to Weather, Climate, and the Environment*, 9th edn, Brooks/Cole Pub Co., 243 pp. Classic general meteorology book, easy to understand with nice clear diagrams.

ascom, W., 1980. *Waves and Beaches: The Dynamics of the Ocean Surface*, Anchor Press, 366 pp. A classic, easy-to-read book, highly descriptive.

lc.Ilveen, R., 2010. *Fundamentals of Weather and Climate*, 2nd edn, Oxford University Press, 448 pp. Another good basic meteorology book, with a few semi-advanced concepts tucked away in appendices at the end of each chapter.

pen University, 2000. *Waves, Tides and Shallow-Water Processes*, 2nd edn, Butterworth-Heinemann, 228 pp. One of a series of excellent introductory books on basic physical oceanography, with plenty of diagrams and simple examples.

inet, P., 2009. *Invitation to Oceanography*, 5th edn, Jones & Bartlett, 626 pp. Another very good introductory oceanography book, covering a wide range of subjects and totally suitable for anyone with no previous knowledge.

verdrup, K. and Armbrust, V., 2008. *An Introduction to the World's Oceans*, 10th edn, McGraw Hill, 528 pp. Excellent introductory oceanography book, full of diagrams, photos and clear, simple explanations.

Textbooks: slightly mathematical

olthuijsen, L., 2010. *Waves in Oceanic and Coastal Waters*, Cambridge University Press, 404 pp. Contains some very useful material on ocean wave generation, propagation and shoaling, with quite a lot of maths but good verbal descriptions and some neat diagrams.

omar, P., 1998. *Beach Processes and Sedimentation*, 2nd edn, Prentice Hall, 430 pp. One of the most wide-ranging reference books available on waves, beaches and coastal processes, with plenty of descriptive text and a few graphs and equations.

lasselink, G. and Hughes, M., 2011. *Introduction to Coastal Processes and Geomorphology*, 2nd edn, Hodder Education, 432 pp. Modern textbook dealing with waves, tides and other processes. Includes a few simple equations.

ugh, D., 2004. *Changing Sea Levels: Effects of Tides, Weather and Climate*. Cambridge University Press, 280 pp. Comprehensive treatment of the tides and everything to do with them, but with just enough descriptive text to make it interesting without having to suffer the maths.

Textbooks: considerably mathematical

insman, B., 2002. *Wind Waves: Their Generation and Propagation on the Ocean Surface*, 2nd edn, Prentice Hall, 704 pp. Originally published in 1965, this book contains the most complete description you could imagine of all the fundamental theories behind waves.

omen, G. K., Cavaleri, L., Donelan, M., Hassleman, K., Hassleman, S. and Janssen, P., 1996. *Dynamics and Modelling of Ocean Waves*, Cambridge University Press, 556 pp. Definitive guide to wave generation, propagation and forecasting by the team responsible for the WAM – a landmark wave-forecasting model.

toker, J. J., 1992. *Water Waves: The Mathematical Theory with Applications*, 2nd edn, Wiley Classics, 600 pp. Another very comprehensive but highly mathematical treatment of waves, first published in 1957.

Glossary

Note: The definitions given here are not always ‘official’ or general definitions; instead, I have tried to stick to more specific definitions useful for understanding this book. For example, the word *bandwidth* could be applied to many branches of physics or radio science, but here we are only interested in its meaning as applied to ocean waves.

A

ccretion, **accreting**. When a beach is accreting, the sand is moving onshore, usually in small-wave conditions where it moves onshore slowly; however, sudden accretion can also occur in storm conditions.

ccretionary beach rips. Relatively stable rips that occur in small-wave conditions when the sand is gradually moving onshore.

dvection. Movement of a fluid in a horizontal direction.

iry wave theory. A set of equations used to describe wave motion, whereby the **wave height** is assumed to be insignificant compared with the water depth and **wavelength**. Named after George Biddell Airy.

longshore drift. A current that runs along the shore and is produced by the broken waves; occurs when the waves approach the shore at an oblique angle, such as in a pointbreak.

mphidrome, **amphidromic point**. Imaginary point on the sea surface around which the tides circulate; the further away from an amphidrome, the larger the **tidal range**.

mplitude. The amplitude of a wave is the vertical displacement of the water surface from its resting position.

nalysis chart. A weather chart based on real data collected a short time before the chart is published, as opposed to a **forecast chart**, which is based on the output of a simulation model.

nticyclone. An area of high pressure, the air around which circulates clockwise in the Northern hemisphere and anticlockwise in the Southern hemisphere.

stronomical tide. The tidal movements that can be explained by the movements of the celestial bodies only, but not other factors such as local weather conditions.

B

andwidth. The distribution of wave energy over different periods; the energy is spread over a large range of periods with a broad bandwidth, and a smaller range of periods with a narrow bandwidth.

aroclinic instability. When a local spatial change in pressure causes a disturbance in the atmosphere.

arycentre. The axis about which the Earth-Moon system rotates, which is an imaginary point inside the Earth, just near the surface.

athymetric defocusing. The bending outwards of a wave-front, due to the middle propagating faster than the end(s). The same as **convex refraction**.

athymetric focusing. The bending inwards of a wave-front, due to the middle propagating more slowly than the end(s). The same as **concave refraction**.

athymetrically controlled. A bathymetrically controlled rip is one whose characteristics depend on the shape of the sea-floor rather than on the prevailing wave conditions.

athymetry. The shape of the sea-floor, resulting in different water depths at different positions.

eaufort scale. A scale of numbers from zero to 12 (although higher numbers are used in some parts of the Far East) that relates wind speed to observed conditions at sea or on land, devised in 1805 by Francis Beaufort.

ergen school. A group of Scandinavian meteorologists from Bergen in Norway, who were the first people to seriously study the formation of a low-pressure system.

ore. A scientific term for broken waves or lines of whitewater.

reaking profile. The side view or profile of a breaking wave, which varies in shape according to the shape of the sea-floor and the incoming **swell** characteristics.

C

ape Doctor. A strong south-easterly wind that blows around Cape Town, South Africa, typically in summer. The wind is notorious for causing strong **upwelling**, which considerably lowers the water temperature.

ircumferential dispersion. The spreading of **swell** over a progressively wider area as it propagates away from the **storm centre**.

oastal upwelling, see **upwelling**.

old front. Imaginary line marking the transition between warm and cold air in a low-pressure system. Signified on the weather chart by a thick line with triangles on it.

ommon centre of mass. The natural balance point between two objects; in the case of the Earth and the Moon it is also the axis about which they rotate (see **barycentre**).

oncave refraction. The bending inwards of a wave-front, due to the middle propagating more slowly than the end(s). The same as **bathymetric focusing**.

onjunction. When the Moon and Sun are lined up and lying on the same side of the Earth, resulting in a new Moon and **spring tides**.

onstructive interference. When two wave-trains interact and the crest of one wave coincides with the crest of another. The result is additive – i.e., a larger wave is produced.

onvection. Movement of a fluid which carries heat from one part of a system to another in a vertical direction.

onvex refraction. The bending outwards of a wave-front, due to the middle propagating faster that the end(s). The same as **bathymetric defocusing**.

oriolis force. An 'apparent force', due to the rotation of the Earth, which acts perpendicular to a body's motion, causing it to turn to the right in the Northern hemisphere and to the left in the Southern hemisphere. Named after Gaspard Gustave de Coriolis.

yclone. An area of **low pressure** (including tropical storms) around which the air circulates anticlockwise in the Northern hemisphere and clockwise in the Southern hemisphere.

yclonic. Anticlockwise in the Northern hemisphere and clockwise in the Southern hemisphere.

D

estructive interference. When two wave-trains interact and the crest of one wave coincides with the trough of another. The result is subtractive – i.e., the two waves cancel each other out.

irectional spectrum. The distribution of wave energy over period and direction at a single point on the ocean surface. It is often shown as a 'polar' plot with colour contours.

irectional spread. The spread of the wave direction around the direction of maximum energy. A wide directional spread means that the waves are arriving from a lot of different directions at the same time.

ispersion. The spreading out of a **swell** as it propagates away from the **storm centre** (see **circumferential dispersion** and **radial dispersion**).

uration. The length of time for which the wind blows over a specific area of the sea surface to generate waves.

uration limited. When the wave growth is limited by the duration – i.e., the **wave height** depends only on the wind speed and the duration, but not the **fetch**.

ynamic fetch. When the **swell**-producing wind is moving at approximately the same speed as the swell itself, extra energy is transferred into the waves, making them grow higher than normal.

ynamic Tidal Theory. A theory of tides developed by Pierre-Simon Laplace in the early nineteenth century, which superseded the

Equilibrium Tidal Theory developed earlier by Isaac Newton.

E

mpirical, empirical model. Based on the relationship between observed data, but not necessarily on the laws of physics.

quilibrium Tidal Theory. The first proper explanation of the tides based on the movements of the celestial bodies, developed by Isaac Newton in the seventeenth century.

rosion. When a beach erodes, the sediment is moved away from the shore, typically over a short period of time in large wave or storm conditions.

rosionary beach rips. A type of rip that occurs in large-wave conditions, often associated with sediment moving offshore. Their characteristics are controlled by the prevailing wave conditions, not the existing shape of the sea-floor.

xplosive cyclogenesis. Rapid deepening of a **low pressure**. More specifically, when the pressure decreases by more than 24 millibars in 24 hours.

xtra-tropical cyclone. An atmospheric disturbance found at latitudes between about 30° and 70°, resulting in a large mass of cyclonically rotating surface air (clockwise in the Southern hemisphere and anticlockwise in the Northern hemisphere). The most important feature for producing surf. Same as **mid-latitude depression** or **low pressure**.

F

eeder current. The water flowing along the shore feeding into a seaward-flowing rip current.

etch. The length of an area of sea surface over which the wind blows continuously to generate waves.

etch-limited. When the wave growth is limited by the fetch – i.e., the **wave height** depends only on the wind speed and the fetch, not the duration.

orecast, **forecast chart**. A set of values or a weather chart based on the output of a computer model rather than measured data.

requency. The wave frequency is the number of waves that pass a fixed point in a given time interval. The frequency in Hertz is the number of waves that pass a fixed point per second.

ully arisen sea. When the waves cannot grow any higher for a particular wind speed, no matter how long it blows for or over what distance it blows. In a fully arisen sea the **wave height** depends only on the wind speed.

G

ravitational force. The attraction due to gravity between two bodies. Within the context of tides, it is the force that stops the **inertial force** flinging the Earth and Moon apart.

reat-circle route. The shortest route between two points on a sphere, such as the Earth.

roup speed. The speed at which the energy of a **swell** is carried. In deep water, the group speed is half the speed of the individual waves inside the group.

rouping. The formation of waves into sets or groups, due to the interaction between two or many different wave-trains consisting of waves of different periods and directions.

roupiness. The characteristics of the waves in terms of how they are grouped – e.g., the number and variability of waves in the group, and the time between the arrival of successive groups.

H

eight, see **wave height**.

ydraulically controlled. A hydraulically controlled rip is one whose characteristics are controlled by the prevailing wave conditions rather than the shape of the sea-floor.

ysteresis. A delayed reaction to the input of energy to a system; for example, in places where the sea temperature varies with the seasons, the maximum sea temperature often occurs a month or more after the maximum input from the Sun (mid-summer).

I

ıertial force. The force which tries to keep a body travelling along its original trajectory. In the context of tides, it is the force that stops the **gravitational force** pulling the Earth and Moon together.

ıterpolate, **interpolation**. The estimation of the value of some parameter at a point, based on its known value at two or more surrounding points. Useful for plotting contour charts of things like **wave height**.

ibarren number. A number used to distinguish between different types of breaking waves. Calculated from **wave height**, wave **period** and beach slope.

ıobar chart. A weather chart showing how the surface atmospheric pressure is distributed over a certain area, such as the Atlantic Ocean. The same as **surface pressure chart**.

ıobar. A line of equal pressure, found on an **isobar chart**. The closer together the isobars, the stronger the wind.

J

ıt stream. A strong circumpolar westerly air stream at mid-latitudes, at heights of between 5,000 and 10,000 metres. The jet stream is highly influential in the formation and movement of low-pressure systems.

L

ınd breeze. A light offshore wind that sometimes blows from late evening to early morning, caused by warm air sinking over the land and being forced seaward as the land cools.

near superposition. When two or more wave-trains coincide, the vertical positions of the sea surface corresponding to each wave train simply add together to make the resultant wave train.

ow pressure. An atmospheric disturbance found at latitudes between about 30° and 70°, resulting in a large mass of cyclonically rotating surface air (clockwise in the Southern hemisphere and anticlockwise in the Northern hemisphere). The most important feature for producing surf. The same as **mid-latitude depression** or **extra-tropical cyclone**.

M

ıean sea level. The average vertical position of the water surface.

ıegarips. Large, powerful rips that occur in large wave conditions, usually when the **wave height** is above three metres.

ıid-latitude depression. An atmospheric disturbance found at latitudes between about 30° and 70°, resulting in a large mass of cyclonically rotating surface air (clockwise in the Southern hemisphere and anticlockwise in the Northern hemisphere). The most important feature for producing surf. The same as **low pressure** or **extra-tropical cyclone**.

ıonochromatic. A monochromatic wave train has only one period, which is not normally observed in the real ocean because the sea surface always contains waves of different periods mixed together.

N

AO. The North Atlantic Oscillation: a climatic cycle in the North Atlantic which switches between 'mobile' and 'blocking' states, the former characterized by a continuous supply of storms, and the latter by a large, static **anticyclone**, which blocks the storms.

AO index. A number used to estimate whether the NAO is in one state or the other, usually based on the difference in atmospheric pressure between two points – for example, the Azores and Iceland.

atural resonant frequency. The natural frequency at which a system will oscillate, which depends on the characteristics of the system, such as the length of a guitar string or the width of a lake.

eap tides. Tides with the smallest **tidal range** in the **spring-neap cycle**.

egative feedback. When part of the output of a system is fed back into the input at the opposite phase to the input, resulting in an attenuation of the signal.

on-linear transfer or **non-linear wave-wave interactions**. The transfer of energy between waves of different **wavelengths** in a growing sea. One of the most important processes taken into account in wave-prediction models.

orth Atlantic Oscillation, see **NAO**.

O

bliquity of the ecliptic. The angle of 23.5° between the axis of the Earth's rotation around itself and the axis of the Earth's rotation around the Sun.

ccluded front. When a low-pressure system is weakening, the warm and **cold front**s join together, forming an occluded front.

pposition. When the Moon and Sun are lined up and lying on opposite sides of the Earth, resulting in a full Moon and **spring tides**.

P

eak direction. The wave direction at which there is most energy.

eak period. The wave period at which there is most energy.

eriod. The wave period is the time taken for an entire wave to pass a fixed point.

hase speed. The speed of each individual wave.

lunging breaker. The technical term for a tubing wave.

olar front. An imaginary line at mid-latitudes where cold Equator-ward air, originating from high latitudes, meets warm poleward air from low latitudes. The polar front is an example of a place where **low pressures** are likely to form.

ositive feedback. When part of the output of a system is fed back into the input at the same phase as the input, resulting in an amplification of the signal.

ressure gradient. The change in atmospheric pressure over a given horizontal distance. The stronger the pressure gradient, the stronger the wind.

ulsating rips. Rips that vary in intensity over time, usually associated with the wave groups.

Q

uadrature. When the Moon and Sun are lying at right-angles to each other relative to the Earth, resulting in a half Moon and **neap tides**.

R

adial dispersion. In a propagating **swell**, because longer waves travel faster than shorter ones, the longer ones progressively out-distance the shorter ones.

efraction. The bending of a wave as it propagates over different depths.

esonance. When the oscillations of a system match the frequency of energy input, so that the oscillations are maintained or increase, such as bouncing a ball or pushing someone on a swing.

p head. The point at which a rip starts to dissipate, usually some distance beyond the wave-breaking zone.

oaring forties. A band of westerly winds that blow around the globe at a latitude of approximately 40° south. The band is wider and the winds are stronger in winter.

otational rips. Rips that double back towards the breaking waves, forming a closed circuit. Often seen at big-wave spots.

S

ea breeze. An onshore wind that occurs in some places from late morning onwards, caused by air rushing in from the sea to fill the gap produced by warm air rising over the land as it warms up.

emi-diurnal. Semi-diurnal tides are tides that have two highs and two lows a day.

et-up. When there are large waves or a strong onshore wind, water is pushed up on to the shore, causing the average water surface to slope upwards towards the land.

hoaling. When waves start to propagate over progressively shallower water.

idereal month. The time taken for the Earth-Moon system to rotate around its common axis (the **barycentre**), namely 27.3 days.

ignificant wave height. A wave height originally defined to be approximately equal to the most common height estimated by an experienced observer. Significant wave height is quantified as the average height of the highest third of all waves observed over a short period of time.

pecific heat capacity. The amount of heat energy required to raise the temperature of one kilogram of a substance one degree Celsius.

pectrum. The amount of wave energy at each different period. Can be displayed as a two-dimensional plot.

pilling breaker. Technical term for a mushy, slow-breaking wave with no tube.

pring-neap cycle. The tides go through a cycle of about fifteen days where the **tidal range** increases, decreases and then increases again; the cycle is governed by the position of the Moon relative to the positions of the Earth and Sun.

prings, **spring tides**. Tides with the smallest **tidal range** in the **spring-neap cycle**.

tanding wave. A wave that moves back and forth in a water body bounded at both ends, such as a swimming pool or lake, but also includes large bodies such as the Atlantic Ocean.

tatic pressure effect. The effect of the atmospheric pressure on the level of water. For every millibar difference in atmospheric pressure the level of water changes one centimetre.

tatistical highs and lows. Imaginary high- and low-pressure systems derived from the average pressure over a given time, usually the whole summer or whole winter.

teepness. The wave steepness is the **height** of a wave divided by its **wavelength**.

torm centre. The area where the wind is transferring energy into the sea to produce waves.

torm surge. The effect of a large **swell** and/or onshore wind combined with a low atmospheric pressure, which causes the water level at the shore to rise, perhaps inundating man-made structures.

treamlines. In fluid dynamics, a streamline is the path traced out by a particle as it moves with the flow. Streamlines are useful for visualizing the flow of air around objects such as aeroplane wings, or ocean waves.

urface pressure chart. A weather chart showing how the surface atmospheric pressure is distributed over a certain area, such as the Atlantic Ocean. The same as an **isobar chart**.

urging breaker. A wave that doesn't actually break, but just washes up and down the beach. Most commonly found on very steep beaches.

well. Waves that are propagating away from the **storm centre** 'on their own', without any further input of energy from the wind.

well front. On some wave prediction charts, a line marking a discontinuity in wave period, signifying the arrival of long-period waves at the beginning of a **swell**.

T

iermally stratified. If the ocean is thermally stratified, warmer, lighter water sits on top of cooler, heavier water.

dal bulges. In Newton's model of the tides, the water on the Earth has two bulges, one either side, caused by the motion of the Earth-Moon system and the **gravitational force** between them.

dal range. The difference between the water level at low tide and the water level at high tide.

dal wave. The concept of the tide behaving as an extremely long wave circling the planet from east to west, which allows several features of its behaviour to be explained.

de-raising forces. The forces originating outside the Earth that provide the initial input to the tides.

opographically constrained. Topographically constrained rips are rips whose characteristics are controlled by some physical feature sticking out above the water surface, such as a headland or an island, rather than the shape of the sea-floor.

ade-winds. Winds that blow on the Equator-ward flank of high pressures, approximately from the north-east in the Northern hemisphere and from the south-east in the Southern hemisphere.

U

nidirectional. A unidirectional wave-train is travelling in one direction only, which is not normally observed in the real ocean, as the surface always contains waves of different directions mixed together.

pwelling. A phenomenon seen on the western sides of continents, whereby the **trade-winds** constantly blow warm surface water away from the coast, allowing cold water to rise up from underneath.

W

arm front. Imaginary line marking the transition between cold and warm air in a low-pressure system. Signified on the weather chart by a thick line with round blobs on it.

arm sector. The section between the **warm front** and **cold front** in a low-pressure system. This is where the wind normally blows strongest in a straight line.

avebuoy. A floating instrument station which takes real-time measurements of a variety of parameters, including **wave height**.

ave height. The vertical distance between the trough and the crest of a wave.

avelength. The horizontal distance between a particular part of a wave (say, the crest) and the same part of the next wave; or simply the horizontal distance between the beginning and the end of a wave.

/hitecapping. When the wind is strong enough to force the tops of existing waves to break, thus removing energy from the waves.

/ind barb symbols. On wind charts, symbols depicting wind strength and direction at a specific point on the surface.

/indsea. A mixed-up sea containing waves of many different heights, **wavelengths** and directions. In a windsea, waves are still being generated by the wind, and have not started propagating away as **swell**.

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Credits

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